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AN ASSESSMENT OF MORTON'S MODEL FOR ESTIMATING AREAL EVAPOTRANSPIRATION

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ABSTRACT

The complementary relationship areal evapotranspiration model (CRAE) proposed by Morton (1983) is evaluated and assessed with respect to:

(a) theory and formulation,

(b) the procedures and methods used to describe the model components,

(c) the sensitivity of model output to input variables, and

(d) the performance of the model for estimating seasonal and daily evaporation/evapotranspiration amounts, from the disappearance of the seasonal snowcover to harvest, in a Prairie environment.

The three major weaknesses of the CRAE model are its' radiation algorithm, the lack of a routine for estimating the net energy available for evaporation/evapotranspiration during the period of soil thawing and the assumption that the vapour transfer coefficient is independent of wind speed.

CRAE appears to provide better estimates of the "Regional" evapotranspiration during periods of high evapotranspiration demand. It is, however, sensitive to abrupt changes in the dominant weather in the calculation period and in the case of these events the accuracy and representativeness of evapotranspiration estimate is affected by the length of the averaging period used.

The CRAE model presented by Morton (1983) appears to be an appropriate framework on which to base the development of a system which is capable of providing short-term (daily) estimates of "areal" evaporation/evapotranspiration from field-sized areas having specific soil, surface cover and climatic conditions. It is recommended that:

(1) For net all-wave radiation,

(a) The routine for the transmittancy of clouds to extra-terrestrial radiation be modified/changed to permit estimation of the direct and diffuse components of the incident global flux;

(b) The procedure for estimating snow surface albedos be replaced and the routine be modified to allow the input of site-specific albedo values;

(c) The algorithm of net long-wave flux be revised, in particular the downward-directed component and the effects of ambient water vapour pressure on the net flux;

(d) The monthly structure be replaced with a flexible framework to permit estimation of components for short-time (daily) periods.

(2) An algorithm be developed to account for the effects of the amount of soil heat used for the phase change and soil warming in thawing soils on the net energy available for evaporation/evapotranspiration.

(3) An algorithm be incorporated to account for the effects of wind speed on the vapour transfer coefficient.

TABLE OF CONTENTS

LIST OF FIGURES ii

LIST OF TABLES v

EXECUTIVE SUMMARY vi

1. INTRODUCTION 1

2. EVALUATION OF THE THEORY AND VALIDITY OF APPLIED ASSUMPTIONS . . . 3
   2.1 The Complementary Relationship 3

3. APPRAISAL OF THE CONCEPTS OF POTENTIAL AND WET-ENVIRONMENT EVAPOTRANSPIRATION RATES USED BY THE MODEL. 7
   3.1 The Concept Of Potential Evaporation. 7
   3.2 Constraints Placed on the Definition of Potential Evapotranspiration by Bouchet's Development. 12
   3.3 Potential and Wet-environment Evapotranspiration Rates Used in the CRAE Model. 13
      3.3.1 Morton's definition and application of potential evapotranspiration. 13
      3.3.2 Morton's definition and application of wet-environment evapotranspiration. 14

4. EVALUATION OF EXPRESSIONS. 16
   4.1 Evaluation of Expressions for Calculating Net Radiation. . . 16
      4.1.1 Net short-wave radiation 17
         4.1.1.1 Incident short-wave radiation. 17
         4.1.1.2 Albedo. 19
      4.1.2 Long-wave radiation. 24
      4.1.3 Net all-wave radiation 26
      4.1.4 Summary. 29
   4.2 Evaluation of the Vapor Transfer Coefficient. 29

5. SENSITIVITY ANALYSIS 34
   5.1 Temperature. 34
   5.2 Humidity. 35
   5.3 Insolation and Radiation. 37
   5.4 Averaging Period. 40

6. COMPARISON OF MODEL ESTIMATES OF AREAL EVAPOTRANSPIRATION WITH ESTIMATES OBTAINED FROM MONITORED SOIL MOISTURE BALANCES. 42

7. SUMMARY AND CONCLUSIONS. 55

8. RECOMMENDATIONS 58

9. REFERENCES 59

LIST OF FIGURES

Figure 4.1 CRAE versus measured daily incident short-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78. 20

Figure 4.2 CRAE versus measured daily incident short-wave radiation fluxes at Bad Lake for Postmelt and Summer periods, 1981-83. . . . 20

Figure 4.3 Ratio of CRAE to measured daily incident short-wave radiation against sunshine ratio for periods with snow cover, Bad Lake, 1972-78. 21

Figure 4.4 Ratio of CRAE to measured daily incident short-wave radiation against sunshine ratio for Postmelt and Summer periods, Bad Lake, 1981-83. 21

Figure 4.5 CRAE versus measured daily albedos at Bad Lake during periods with snow cover, 1972-78. 23

Figure 4.6 CRAE versus measured daily albedos at Bad Lake during Postmelt and Summer periods, 1981-83. 23

Figure 4.7 CRAE versus measured daily net short-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83. 25

Figure 4.8 CRAE versus measured daily net long-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78. 27

Figure 4.9 CRAE versus measured daily net long-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83. 27

Figure 4.10 CRAE versus measured daily net all-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78. 28

Figure 4.11 CRAE versus measured daily net all-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83. 28

Figure 4.12 Vapor transfer coefficient plotted with wind speed for various vegetated surfaces and for a Class A evaporation pan. . . . 32

Figure 5.1 Effect of a systematic error of ∓0.5°C in temperature on the CRAE estimate of evapotranspiration at Saskatoon, 1981. 37

Figure 5.2 Effect of a systematic error of ∓5% in relative humidity on the CRAE estimate of evapotranspiration at Saskatoon, 1981. 38

Figure 5.3 Effect of a systematic error of ∓0.20 h in sunshine duration on the CRAE estimate of evapotranspiration at Saskatoon, 1981. . . 39

Figure 5.4 Effect of a systematic error of ∓1.5 MJ/m^2-d in calculated net all-wave radiation on the CRAE estimate of evapotranspiration at Saskatoon, 1981. 40

Figure 6.1 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1981. 43

Figure 6.2 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1982. 44

Figure 6.3 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1983. 44

Figure 6.4 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1981. 45

Figure 6.5 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1982. 45

Figure 6.6 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1983. 46

Figure 6.7 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1984. 46

Figure 6.8 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration for March, April and May at Bad Lake (1981-83) and Saskatoon (1981-84). 47

Figure 6.9 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration from wheat for June, July and August at Bad Lake (1981-83) and Saskatoon (1981-84) 47

Figure 6.10 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration from fallow for June, July and August at Bad Lake (1981-83) and Saskatoon (1981-84). 48

Figure 6.11 CRAE and measured regional evapotranspiration at Bad Lake, 1981. 49

Figure 6.12 CRAE and measured regional evapotranspiration at Bad Lake, 1982. 50

Figure 6.13 CRAE and measured regional evapotranspiration at Bad Lake, 1983. 50

Figure 6.14 CRAE and measured regional evapotranspiration at Saskatoon, 1981. 51

Figure 6.15 CRAE and measured regional evapotranspiration at Saskatoon, 1982. 51

Figure 6.16 CRAE and measured regional evapotranspiration at Saskatoon, 1983. 52

Figure 6.17 CRAE and measured regional evapotranspiration at Saskatoon, 1984. 52

Figure 6.18 CRAE using a wind function for vapor transfer, and measured regional evapotranspiration at Bad Lake, 1981. 54

Figure 6.19 Regional estimates by CRAE (without a wind correction) plotted against measured evapotranspiration for June, July and August at Bad Lake (1981, 1982) and Saskatoon (1982, 1983). 54

Figure 6.20 Regional estimates of evapotranspiration by CRAE (with wind correction) plotted against measured evapotranspiration for June, July and August at Bad Lake (1981, 1982) and Saskatoon (1982, 1983). 54

LIST OF TABLES

Table 3.1 Summary of definitions describing "measured" potential evapotranspiration parameters (after Perrier, 1977). 10

Table 3.2 Summary of definitions describing "calculated" potential evaporation parameters. (from Granger, 1989a) 11

Table 4.1 Standard deviation of the 10-m mean wind speed as a function of measurement period for Saskatoon (SRC) (1978 - 1987). 33

Table 5.1 Thirty-day evapotranspiration totals (mm) in 1981 and 1983 at Bad Lake, as calculated by the CRAE model using averaging periods of 5-, 10-, and 30-days. 40

EXECUTIVE SUMMARY

The complementary relationship areal evapotranspiration model (CRAE) proposed by Morton (1983) is evaluated and assessed in respect to: (a) theory and formulation, (b) the procedures and methods used to describe model components, (c) the sensitivity of model output to input variables and (d) the perfomance of the model for estimating seasonal and daily evaporation/evapotranspiration amounts over the period extending from the disappearance of the seasonal snowcover up to the time of harvest of annual cereal crops in a Prairie environment. Morton's model is based on the relationship developed by Bouchet (1963) who demonstrated that there is a negative or complementary response of potential evapotranspiration to changes in the availability of water for areal evapotranspiration and can be represented by:

$$E_T + E_{TP} = 2E_{TW},$$

in which E_T = actual evapotranspiration, E_{TP} = potential evapotranspiration and E_{TW} = wet environment evapotranspiration. Morton suggests that the complementary relationship cannot be verified with currently available theoretical knowledge and empirical evidence. However, he assures the validity of Bouchet's relationship by using the complementary relationship itself as the basis for the definition of E_{TW} and establishes the wet-environment evapotranspiration rate equal to one-half the dry-environment potential evapotranspiration rate (as he has defined it).

The approaches used by CRAE (for example: empirical relationships obtained from monthly data; approximate solutions for albedo; neglect of the effects of wind speed and surface roughness; calibrations against large-scale river basin water budget estimates) to estimate different components of the parameters of the complementary relationship are such that the model loses its predictive capabilities for short-term simulations of special case environments, such as for snow cover or for single, field-sized areas. The three major weaknesses of the CRAE model are its' radiation algorithm, the lack of a routine for estimating the net energy available for evaporation/evapotranspiration during

the period of soil thawing and the assumption that the vapor transfer coefficient is independent of wind speed.

(1) The radiation model used by CRAE and the monthly structure of the calculations result in significant errors when shorter time periods are used. It is demonstrated that there are also significant errors in the estimation of the transmittancy of clouds to extra-terrestrial radiation, surface albedo and net long-wave radiation. The resultant error introduced in net "all-wave" radiation can cause appreciable error in the estimate of seasonal evapotranspiration. Also, CRAE requires accurate humidity data and the accuracies generally associated with the most commonly-used methods of humidity measurement (lithium chloride dewcell, ∓3% RH; hygrothermograph, ∓5%) are inadequate.

(2) The assumption that the soil heat flux and storage terms are negligible restricts the application of CRAE to those periods when soil is not freezing or thawing. In the case of soil thawing an appreciable amount of available energy may be apportioned to the soil heat flux term which supplies the energy required for the phase change and soil warming. It is demonstrated that CRAE consistently overestimates actual evapotranspiration during the early spring months of soil thaw.

(3) The assumption that the vapor transfer coefficient is independent of wind speed is not appropriate. It is demonstrated that neglecting the effects of wind speed can result in significant errors, even when monthly periods are used. As well, neglecting the effect of wind speed means that the effects of surface roughness are also ignored; which limits estimates to a "Regional" scale and prevents the model from simulating the effects on evapotranspiration of natural or man-made changes to the environment.

CRAE appears to provide better estimates of the "Regional" evapotranspiration during periods of high evapotranspiration demand. It is, however, sensitive to abrupt changes in the dominant weather in the calculation period and in the case of these events the accuarcy and representativeness of evapotranspiration estimate is affected by the length of the averaging period used.

The CRAE model presented by Morton (1983) appears to be an appropriate framework on which to base the development of a system which is capable of providing short-term (daily) estimates of "areal" evaporation/evapotranspiration from field-sized areas having specific soil, surface cover and climatic conditions. To achieve this objective it is recommended that:

(1) For Net All-wave Radiation,

   (a) The routine to account for the transmittancy of clouds to extra-terrestrial radiation be modified or changed to permit the estimation of direct and diffuse components of the incident global flux.

   (b) The procedure for estimating snow surface albedos be replaced and the routine be modified to allow the input of site-specific albedo values.

   (c) The algorithm of net long-wave flux, be revised, in particular as regards to the downward-directed component and the effects of ambient water vapor pressure on the net flux.

   (d) The monthly structure be replaced with a more flexible framework so as to permit the estimation of components for short-time (daily) periods.

(2) An algorithm be developed to account for the effects of the amount of soil heat used for the phase change and soil warming in thawing soils on the net energy available for evaporation/evapotranspiration.

(3) An algorithm be incorporated to account for the effects of wind speed on the vapor transfer coefficient.

Several of these modifications will also improve the performance of the model for estimating evapotranspiration over integrated spatial and temporal scales.

1. INTRODUCTION

Morton (1983) presents a Complementary Relationship Areal Evapotranspiration model based on a relationship developed by Bouchet (1963) who showed there is a negative, and possibly complementary, response of potential evapotranspiration to changes in the availability of moisture at the evaporating surface. Bouchet examined the changes in the energy balance of a region as the surface moisture changed and, although the energy balance represents a sound theoretical framework, the development of his hypothesis involved a certain amount of speculation.

In the presentation of his CRAE model, Morton stresses its applicability to a wide variety of environments, and suggests that there is no need for locally-optimized calibration coefficients. However, he does impose the limitations that it cannot be used for short time intervals or near sharp environmental discontinuities. In other words, he recommends it's use for integrated spatial and temporal scales. There are, however, many situations which require reliable estimates of evapotranspiration for short time periods and for relatively-small areas, for example areal units of field size.

In this report Morton's Complementary Relationship Areal Evapotranspiration model is examined in respect to theory and applied assumptions, the validity of the algorithms used to describe different terms and the applicability of the model for estimating evapotranspiration on reduced spatial or temporal scales from those which it was developed. The assessment covers the following points:

1. Evaluation of the theory and validity of applied assumptions used in CRAE and an assessment of the limitations imposed on model performance by the applied assumptions.
2. Appraisal of the "concepts" of potential and wet environment evaporation used by the model.
3. Evaluation of expressions for calculating net radiation based on measured fluxes.
4. Sensitivity analysis; response of model to different inputs.

5. Comparison of CRAE estimates of areal evapotranspiration with estimates obtained from monitored soil moisture changes.

6. Recommendations.

The discussion herein concerns the version of CRAE presented by Morton (1983). In order to allow the reader to compare the works, the nomenclature used by Morton (1983) has been adopted.

2. EVALUATION OF THE THEORY AND VALIDITY OF APPLIED ASSUMPTIONS

2.1 The Complementary Relationship

Morton's Complementary Relationship Areal Evapotranspiration model (CRAE) is based on the theory, or the observation, that there exists a negative response of potential evaporation to changes in the amount of water available for areal evaporation. This concept was first introduced by Bouchet (1963) who showed, using the energy balance, that when a region was allowed to dry from a saturated condition the reduction in areal evapotranspiration liberated a certain amount of energy which went to increasing the potential evapotranspiration of the region, as evidenced by an increase in air temperatures and reduced vapor pressures. Bouchet postulated that the increase in potential evapotranspiration could be equivalent to the reduction in regional evapotranspiration, in other words that the relationship could be complementary. The complementary relationship that Bouchet derived states that the sum of the "actual", regional (E) and potential (E_p) evapotranspiration rates is equal to twice the wet environment evapotranspiration rate (E_{pw}), that is:

$$E + E_p = 2E_{pw} \qquad 2.1[8]^1$$

This relationship is the basis of Morton's model.

Bouchet cautioned that the relationship was likely approximate and that because of large-scale exchanges or advection the sum of the actual and potential evapotranspiration rates would be less than twice the wet environment evapotranspiration rate. Morton, however, adopts the equality, and presents a mathematical rationalization to justify his choice.

Models based on the complementary relationship between actual and potential evapotranspiration differ from other currently-accepted relationships (eg. the Versatile Soil Moisture Budget) in several ways:

---

[1]Number in brackets corresponds to equation given by Morton (1983).

(1) Other relationships generally assume that increases in potential evapotranspiration entail increases in actual evapotranspiration whereas the complementary relationship suggests that increases in potential evapotranspiration will be reflected as decreases in actual evapotranspiration.

(2) The complementary relationship suggests that potential evapotranspiration is an effect rather than a cause of regional evapotranspiration.

(3) Conventional models generally relate actual evapotranspiration to a single parameter called potential evapotranspiration, whereas complementary relationship models make use of two evapotranspiration parameters, the potential and wet-environment evapotranspiration rates.

(4) Other models which relate actual to potential evapotranspiration are generally based on regressions or empirical relationships and the choice of the definition for potential evapotranspiration is not critical, as long as there is consistency of definitions between model development and use. The complementary relationship, on the other hand, is based on the responses of the energy balance to changes in surface wetness, and the definitions of the parameters (potential and wet-environment evapotranspiration rates) must be consistent with this approach.

Morton (1983) notes that complementary relationship models have the following limitations:

(1) They require accurate humidity data.

(2) They cannot be used for short time intervals. Morton attributes this limitation to subsurface heat storage changes and to the lag times associated with the change in storage of heat and water vapor in the atmospheric boundary layer after changes in surface conditions or the passage of frontal systems. Fortin and Seguin (1975) analyzed the validity of Bouchet's relationship (Eq. 2.1) for different time scales (1 to 30 days) and confirmed the validity of the relationship even for short periods.

(3) They cannot be used near sharp environmental discontinuities, because of heat and water vapor advection in the lower layers of the atmosphere. One important consideration is that representative measurements of the atmospheric parameters be obtained at a height within the fully developed boundary layer. There is no precise evidence of the space scales required to satisfy this condition. De Vries (1959) noted the influence of advection up to 1 km from the leading edge; Davenport and Hudson (1967) indicate that the effects of advection can decrease to near zero within 300 m; and Millar (1964) suggests that these can be negligible as close as 5 m from the edge. However, some of the formulations adopted by Morton in his CRAE model do affect the imposed scale limitations. The fact that the effect of wind on the vapor transfer coefficient is not considered means that the effects of the surface are also ignored, thus limiting estimates to the regional scale.

(4) They require input from a climatological station whose surroundings are representative of the area of interest. Although this might be seen as a "limitation", it is an accepted "requirement" for most models, whether of the complementary relationship type or not.

(5) They cannot be used to predict the effects of natural or man-made changes because they neither use nor require knowledge of the soil-vegetation system. As indicated above, this limitation may be directly attributed, in the case of Morton's CRAE model, to the fact that the procedures used for parameter evaluation are independent of the surface characteristics (ie. vapor transfer coefficient).

The complementary theory for the formulation of an evaporation model should not in itself impose any limitations on the applicability of the model because the theory is based on the energy balance, which is universally applicable. However, the method does not apply the energy balance directly, but rather makes use of known or inferred responses of the terms of the energy balance to changes in the moisture supply to the region. The method thus has the potential of eliminating some of the drawbacks of the application of an energy balance (calculation of all terms; knowledge of surface parameters) while benefitting from some of the advantages (sound theoretical basis; universal applicability).

The theory of a negative response between actual and potential evaporation (the Bouchet relationship) is an appropriate one and has been shown to be correct. Whether the relationship is truly complementary, that is a change in one will result in an equal but opposite change in the other, has not been proven. The exact form of the relationship between actual, potential and wet-environment evaporation can depend on the time and space scales involved; but more importantly, the relationship will depend on the definitions adopted for the potential and wet-environment evaporation rates. This is particularly true in light of the fact that the method is based on the energy balance equation which has been differentiated with respect to available moisture, and this does impose some conditions on the selection of a definition for the term "potential evaporation".

3. APPRAISAL OF THE CONCEPTS OF POTENTIAL AND WET-ENVIRONMENT EVAPOTRANSPIRATION RATES USED BY THE MODEL.

Bouchet (1963), in his paper describing the relationship, defined potential evaporation as the water which would be lost under the same conditions when the water supply is not a limiting factor; he defined the wet-environment evaporation as the potential evaporation rate when the actual and potential rates were equal, that is, when the surface is saturated.

Bouchet's definition of potential evaporation is somewhat vague in that he does not specify what is meant by "under the same conditions". It is unfortunate that Bouchet did not present, or derive, a more precise definition of potential evaporation for use with his relationship. Those authors who have attempted to apply his relationship (Morton, 1978, 1983; Brutsaert and Stricker, 1979) have thus selected definitions from amongst those in common use. These are many and varied, and there exists no clear consensus since the "concept" of potential evaporation itself has never been clearly or universally defined. Before discussing Morton's definition for potential evaporation, it is appropriate to examine briefly the concept of potential evaporation and the various definitions which are current in the literature, and to examine those constraints placed on the definition of potential evaporation by Bouchet's development.

3.1 The Concept Of Potential Evaporation.

Potential evaporation (or potential evapotranspiration) has been freely used as a hydrological parameter for nearly half a century. The term potential evapotranspiration was introduced by Thornthwaite (1948) in the context of the classification of climate:

> "There is a distinction, then, between the amount of water that actually transpires and evaporates and that which would transpire and evaporate if it were available. When water supply increases, as in a desert irrigation project, evapotranspiration rises to a maximum that depends only on the climate. This we may call "potential evapotranspiration", as distinct from actual evapotranspiration."

Thornthwaite then classed a climate as moist or dry depending on whether precipitation was greater or less than potential evapotranspiration.

Although the concept of potential evapotranspiration was first developed as a climate index, it has been used by agriculturalists, interested in efficient use of water by crops, as an indication of the potential for evapotranspiration, and by hydrologists as a more predictable, or accessible, index from which actual evapotranspiration could be estimated. Indeed, the latter is what motivated Penman (1948), who was

> "seeking an absolute relation between weather elements and open water evaporation, and comparative relations between losses from the soil and losses from open water exposed to the same weather."

Penman's intent may not have been to develop or conceptualize a "potential" evaporation; however, perhaps because what he calculated appeared to be similar to what Thornthwaite had defined, his expression came to be known as the Penman Potential Evaporation, and has since been extensively used for this purpose.

Potential evaporation has become a standard ingredient in the development of new schemes for estimating actual evaporation and the majority of evaporation models estimate actual evaporation as some fraction or function of this "maximum potential evaporation". However there is yet no clear consensus as to the proper definition of this important term. Depending on the situation or the intended use, modifications have frequently been introduced into the Penman equation, or completely different and new definitions have been employed. A survey of the literature reveals there are several different definitions (with variations of each) available for potential evaporation. These vary from a simple statement equating potential evaporation to evaporation from a free water surface, or from a saturated or "well-watered" grass surface, to verbal descriptions of the Penman equation, with and without modifications, and to somewhat elaborate descriptions of conditions under which "potential evaporation" might be measured. These definitions do not all concur, and do not of course all yield identical values.

The various definitions of potential evaporation could be classified according to different schemes. A first distinction could be made between "measured" and "calculated" parameters. The "measured" potential evaporation parameters include those which refer to physically attainable situations (ie. evaporation from an irrigated grass surface), those which refer to results

obtained from measurement devices (e.g. evaporation pans), and as well those "models" used to calculate evaporation from various wet surfaces (ie. variations of the Penman equation for estimating evaporation from an irrigated field). The "calculated" potential evaporation parameters refer to those expressions used to calculate evaporation from hypothetical wet surfaces under prescribed conditions.

Perrier (1977) tabulated a number of definitions of potential evapotranspiration of the "measured" category in common use in agronomy, and suggested appropriate terminology for each. Table 3.1, adapted from Perrier, summarizes those measured parameters which have been used to represent potential evapotranspiration. Perrier (1977) suggested that the following relationships apply:

$$EP_s \geq EP > ETP_o, \text{ and} \qquad 3.1$$

$$EP > ETR_m \geq ETR, \qquad 3.2$$

where ETR is the actual evapotranspiration rate.

More recently, Granger (1989a), in an examination of the concept of potential evaporation, suggested that if the main purpose of the potential evaporation parameter is to serve as an index of, or as an aide for estimating the actual evaporation from a natural non-saturated surface, it is not necessary that it represent a real or attainable situation to be useful; it is more important that it be unique, definable, predictable and related to the actual conditions. By considering the evaporating surface as a system to and from which both energy and mass (water vapor) can be transferred, and by defining the hypothetical evaporation rates which would occur if the surface was brought to saturation under various prescribed conditions, Granger (1989a) developed a series of "calculated" potential evaporation parameters, some of which have been used to represent potential evaporation. These are presented in Table 3.2.

Table 3.1 Summary of definitions describing "measured" potential evapotranspiration parameters (after Perrier, 1977).

EP_*	Theoretical Potential Evaporation	The evaporation from a saturated vegetated cover with complete radiation absorption at the upper surface (a hypothetical situation)
EP	Potential evaporation	The evaporation from a saturated cover. (ie. vegetated cover after a rain or irrigation, free water surface)
ETR_m	Maximum Actual Evapotranspiration	The evapotranspiration from a vegetated cover well supplied with water
ETP	Potential Evapotranspiration	The evapotranspiration from an actively growing, well-watered short grass surface
ETP_C	Climatic Potential Evapotranspiration	Refers to the results of formulae or models used to estimate the above parameters
ETP_O	Limiting Regional Potential Evapotranspiration	An evaporation rate related only to available energy (generally equated to the first term of the Penman expression)
E	Point Evaporation	The evaporation rate as measured with evaporation pans, Piche evaporimeters, atmometers, etc.

Table 3.2 Summary of definitions describing "calculated" potential evaporation parameters. (from Granger, 1989a)

Name	Definition	Calculation	Common usage
EP1	Evaporation rate which would occur if the surface was brought to saturation.	Indeterminate.	
EP2	Evaporation rate which would occur if the surface was brought to saturation and the energy supply to the surface was held constant.	Indeterminate. Regression with solar radiation (Priestley and Taylor, 1972)	wet-environment or equilibrium evaporation. advection-free evaporation.
EP3	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the energy supply to the surface were held constant.	Energy balance, vapor transfer equations.	Penman potential. wet-environment evaporation.
EP4	Evaporation rate which would occur if the surface was brought to saturation and the energy supply and the surface temperature were held constant.	Indeterminate.	not used
EP5	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the surface temperature were held constant.	Vapor transfer equation (if surface temp. is known)	Potential (van Bavel, 1966)

With the exception of EP4, which is an impractical parameter, all other evaporation rates have from time to time been referred to as the "potential evaporation" rate. These different potential evaporation rates and the actual rate, E, are related in the following general manner:

$$EP5 \geq EP3 \geq EP2 \geq EP1 \geq E. \qquad 3.3$$

They also behave in a complementary-like manner in that the difference between the respective evaporation rates is greatest for initially-dry conditions and they are of course equal for initially wet conditions. The same could be said of the inequalities presented by Perrier (1977). However, a truly complementary relationship as described by Bouchet (1963), where the differences between respective evaporation rates are equal, remains to be confirmed. Granger (1989b) derived the relationship between E, EP3 and EP5, and showed that for this case it was neither symmetric nor truly complementary.

It may also be possible to define another parameter in such a way that a complementary relationship relates it to the actual and another "potential" evaporation rate. As will be shown, this is in fact what Morton has done with his definition for the wet-environment evaporation.

This brief survey demonstrates that in fact a large number of different expressions and scenarios have been used to represent "potential evaporation" or "potential evapotranspiration". The question remains as to which one might be the "proper" definition for application in Bouchet's relationship.

3.2 Constraints Placed on the Definition of Potential Evapotranspiration by Bouchet's Development.

Although Bouchet's definition of potential evapotranspiration is vague, it does not necessarily follow that the potential evapotranspiration he referred to is the same as that which is "normally" defined or commonly used. (As was shown, there are a large number of different definitions commonly-used for potential evapotranspiration). Whereas there has been a marked tendency to exclude surface parameters from the definitions of potential evapotranspiration (largely because of their general unavailability), Bouchet, in his discussion, shows that potential evapotranspiration cannot be defined solely as a function

of energy, or climatic considerations independently of the water supply (a surface parameter). This, of course, follows from the approach which he used. He was attempting to make use of the fact that there are changes in the partitioning of energy with changes in available moisture. In essence this amounts to differentiating the energy balance equation with respect to available moisture and examining the behaviour of the terms of the equation. By stating that as the surface dries, the energy not consumed by the evaporation process goes to increasing the potential evapotranspiration, Bouchet is in fact imposing the condition that this potential evapotranspiration be governed largely by the surface conditions of moisture and temperature. Because the Penman equation, which is the most commonly-used expression for the potential, is independent of the actual surface conditions, it would appear that it does not represent the potential evapotranspiration rate required by Bouchet's relationship.

3.3 Potential and Wet-environment Evapotranspiration Rates Used in the CRAE Model.

3.3.1 Morton's definition and application of potential evapotranspiration.

Morton, as others have also done, equates pan evaporation and potential evapotranspiration. He thus presents the following definition (which can be viewed as a description of an ideal evaporation pan):

> Potential evapotranspiration...the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible.

Based on the findings by Kohler et al. (1955), Morton suggests that the Penman equation can provide a good estimate of the evaporation from a class-A evaporation pan if modified to take into account both the difference between vegetative and water surfaces and the sensible heat flux through the wall and bottom of the pan. However, nowhere is it evident or apparent that he provides corrections either to pan data to make them compatible with calculations, or to the calculations to make them comparable to pan data.

Seguin (1975) presents another difficulty with Morton's definition of potential evapotranspiration. He points out correctly that any small evaporimeter will be subjected to significant local advection, and that the

value of the potential evapotranspiration measured will depend on the size of the measurement surface. He suggests that it would have been more appropriate for Morton to define the potential as the limit of measured evapotranspiration as the measurement surface approaches zero. His analysis showed that as this surface was reduced, the measured value increased significantly because of the advection of energy from the surrounding area. Referring to Table 3.2, the measured value thus approached (but did not reach) the value of the parameter EP5.

In consideration of the above, what Morton calculates as potential evapotranspiration is somewhat different from what he has defined. He uses the combined energy balance and vapor transfer equations to calculate potential evapotranspiration but does not include an advection term in the energy balance. The result is an energy balance for a large area, that is, an area sufficiently large that the measurement height of atmospheric parameters is within the developed boundary layer. His method then is simply a variation of the Penman approach. Referring to Table 3.2, what Morton calculates as potential is the parameter EP3, while his definition corresponds (in the limiting case as the measurement area goes to zero) approximately to EP5.

In light of the discussion in Section 2.2, it would also appear that Morton's choice of potential evapotranspiration is not the correct one to use in Bouchet's relationship. Morton's potential (as calculated) corresponds to the Penman evaporation, which is independent of the surface parameters, whereas Bouchet's development requires that the potential evapotranspiration be governed largely by the surface moisture and temperature. However, as will be shown, while Morton does not use a definition consistent with Bouchet's development, he ensures a complementary relationship in his model through his selection of a definition for the wet-environment evapotranspiration.

3.3.2 Morton's definition and application of wet-environment evapotranspiration.

Morton defines wet-environment areal evaporation as

"the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water."

Although this definition is not specific (other than a saturated surface no other conditions have been specified for the system), it is clear from the discussion and the fact that, as did Priestley and Taylor (1972) and others, he estimates it with the use of a regression against net radiation, that what is being referred to is in fact the parameter EP2 (Table 3.2), or Perrier's (1977) ETP_0, the limiting regional potential evapotranspiration (Table 3.1). However, whereas Priestley and Taylor (1972) obtained their regression using measured radiation and evaporation over saturated surfaces, Morton correlates the potential evaporation of a dry environment (such as a desert, where it can be assumed that the actual evaporation is zero) against net radiation and assumes that the complementary relationship holds. He thus calculates the wet-environment evapotranspiration as one-half the dry-environment potential evaporation, which was obtained through regression against net radiation.

By taking this approach, Morton achieves two things. First, rather than actually providing a regression equation for the parameter EP2 (or ETP_0), he has in fact defined a new parameter (which he calls wet-environment evapotranspiration) in such a way that it is equal to one-half the potential evaporation (EP3) of a dry surface, and secondly, he ensures the complementary relationship will hold.

In light of the above discussion, it is fair to say that Morton's CRAE model makes use of a workable relationship, although somewhat different from that of Bouchet. Morton's complementary relationship could be expressed in the following manner:

> "The sum of the actual evapotranspiration and the potential evapotranspiration (as calculated by Morton's variation of the Penman equation) is equal to twice the value of a parameter which is defined as one-half the dry-environment potential evaporation (as calculated above)."

4. EVALUATION OF EXPRESSIONS.

In the CRAE model, areal evapotranspiration is calculated as the difference between twice a "wet-environment evapotranspiration" and a "potential evapotranspiration". The latter is estimated by solving the combined energy balance and vapor transfer equations:

$$E_{TP} = R_T - \lambda_p f_T(T_p - T) \qquad 4.1[10]$$

$$E_{TP} = f_T(v_p - v_D) \qquad 4.2[11]$$

in which E_{TP} is the potential evapotranspiration; R_T is the net radiation; T_p and T are the equilibrium surface and air temperatures, respectively; f_T is the vapor transfer coefficient; v_p is the saturation vapor pressure at T_p; v_D is the saturation vapor pressure at the dewpoint temperature; and λ_p is a function of the atmospheric pressure and the equilibrium temperature.

The wet-environment evapotranspiration is obtained from a regression equation relating it to the net radiation. The major expressions to be evaluated in the CRAE model are thus the net radiation and the vapor transfer coefficient.

4.1 Evaluation of Expressions for Calculating Net Radiation.

In any evaporation model based on energy considerations an accurate estimate of the available energy, primarily the net all-wave radiation input, is important. To evaluate Morton's procedure, the empirical expressions used in calculating the components of the net all-wave radiation flux of the 1983 CRAE model (Eqs. C-1 to C-27, Morton, 1983) were compiled in a FORTRAN computer program, set for the latitude and elevation of Bad Lake and organized to accept daily meteorological data for that station. This enabled comparisons of the measured radiative fluxes, albedos, transmittancy, etc. at the site with those calculated by Morton's procedure. The data were stratified into two groups: (a) Snow-covered - includes Premelt and Melt periods (February, March), and (b) Snow-free - period from the disappearance of the seasonal snowcover to the end

of summer (March through September). The data representing days with snowcover are from 1972-78, and the snow-free data are for 1981-83.

As with virtually all radiation models, Morton's procedure divides the net all-wave radiation into its short- and long-wave components. Each term is examined separately.

4.1.1 Net short-wave radiation

The net short-wave radiation at a surface, Q_{sn}, is the difference between the short-wave incident flux, Q_s, and the reflected flux, Q_r. Morton (1983) calculates Q_{sn} as:

$$Q_{sn} = Q_s(1 - A), \qquad 4.3$$

where A is the albedo. The two variables required to solve the net short-wave flux, Q_{sn}, are the incident flux, Q_s, and the surface albedo, A.

4.1.1.1 Incident short-wave radiation. Morton's procedure for determining Q_s involves calculating the extra-terrestrial short-wave flux, Q_A, moderating this term by a function of the clear-sky transmittancy, τ, to obtain the clear-sky short-wave insolation, Q_o, then moderating this term by a function of the cloud cover.

The equation for the extra-terrestrial flux (Eq. C-11) requires prior calculations of the declination of the sun, the noon and mean angular zenith distances of the sun, the solar angle between noon and sunrise, and the radius vector of the sun. The equations for the declination (C-6) and the radius vector (C-10) rely upon integral month values (ie. 1, 2, 3,...12) which are offset by 15 days to make the results indicative of the mid-month conditions. To use other than monthly inputs a cumbersome conversion is used (C-43) which retains the monthly structure. The errors in this approach are not negligible: for example, they reach 3.6% of the value for declination in February which produces an error of 2.6% in Q_A for that month.

The clear-sky insolation, Q_o, is calculated using a formulation by Brooks (1960) (Morton's Eq. C-21) that is presented in a Climatology course syllabus at the University of California, Davis. Interestingly, this formulation and

reference is not cited in more recent comprehensive publications on atmospheric radiation (Sellers, 1965; de Jong, 1973; Oke, 1978). The data used to develop this relationship are crude and the empirical formulations reflect an understanding of atmospheric radiation current in the 1950's. More efficient methods of calculating Q_o, which approximate the physical processes of light scattering and absorption, are available in the literature (for example see Suckling and Hay, 1976). Brooks (1960) formulation moderates the extra-terrestrial radiation, Q_A, by a function of the clear-sky transmittancy, τ. The calculation of τ is also taken from Brooks (1960) (Morton's Eq. C-19) and attempts to account for the effects of elevation (dust), atmospheric water vapor, and temperature. The calculation of atmospheric water vapor is based on a line fit to mean monthly values of saturation water vapor pressure and temperature collected by the Atmospheric Environment Service for 6 months in the 1970's. The scatter in these data suggest an uncertainty of approximately 18% in the estimation of the atmospheric water vapor; estimates for shorter time periods would show greater variation. Morton also applies a constraint (Eq. C-20a) to prevent the effects of absorption by water vapor from exceeding the total effects of absorption and scattering. That a constraint is necessary suggests that the method for calculating transmittancy may not describe the physical process well.

A typical value of transmittancy for the snowmelt period (ie. period of high transmittancy) by the model is 0.67; the observed value at Bad Lake for this period is of the order of 0.85. In the summer when transmittancy is at its lowest (July), the model calculates a transmittancy of 0.60, while the observed value at Bad Lake is 0.68. The difference of 0.18 in spring and 0.08 in summer will result in errors of 18% and 8% in the respective calculations of Q_o.

The incident short-wave flux is divided by Morton into a portion transmitted through clear-sky and a portion through clouds (Eq. C-22). The incident short-wave transmitted through clear-sky, SQ_o, is a straightforward calculation based on the sunshine ratio, S. The radiation transmitted through clouds is calculated using an empirical relationship based on data collected by Rietveld (1978) for Europe. Morton (1983) assumes a linear relationship between the ratio of incident to extra-terrestrial radiation for days with complete cloud cover and the mean sunshine ratio of the season in which these days

occurred. Such form of relationship suggests that clouds during "sunny seasons" are less opaque than those during "cloudy seasons". The data show considerable scatter and it is anticipated that the curve may be tenuous at best for days of partial cloud cover. To summarize, the approach used by Morton for calculating Q_s is different from the currently-accepted practice of estimating direct and diffuse incident radiation, which applies to both conditions of partial and complete cloud cover.

The daily incident short-wave flux calculated by Morton's formulations using Bad Lake meteorological data is plotted against the measured daily global radiation for days with snowcover in Fig. 4.1, and for snow-free days in Fig. 4.2. For Premelt the mean difference between the measured and calculated flux is 2.88 MJ/m^2-d with a standard deviation of 2.90 MJ/m^2-d. Differences during Melt are somewhat smaller with a mean difference of 2.65 MJ/m^2-d and a standard deviation of 2.66 MJ/m^2-d. For the snow-free period the mean difference between the measured and calculated flux is the smallest, 0.77 MJ/m^2-d, but the standard deviation of 2.66 MJ/m^2-d is of similar magnitude as for the other periods.

Figures 4.1 and 4.2 show large underestimations in the calculated values when Q_s is small, suggesting that the larger errors may be associated with cloudy days. To examine the effect of the sunshine ratio on the difference, the ratio of the calculated to measured fluxes is plotted against the observed sunshine ratio for the snow-covered period in Fig. 4.3 and for snow-free conditions in Fig. 4.4. In both figures a systematic increase in the ratio with increasing sunshine ratio is evident. These trends are consistent with the findings reported in Figs. 4.1 and 4.2 which show underestimations by the model increasing with decreasing flux. The results suggest that Morton's procedure does not handle the transmittancy of clouds to extra-terrestrial radiation in an appropriate manner to provide accurate, reliable estimates of the daily incident flux.

On clear days with high fluxes Morton's procedure generally underestimates Q_s during Melt and overestimates the flux during summer.

4.1.1.2 Albedo. Once the incident short-wave radiation is obtained, the surface albedo is required to calculate the net short-wave radiation at the surface. Morton's calculation of the surface albedo involves a series of empirical approximations of special-case albedos before the mean surface albedo

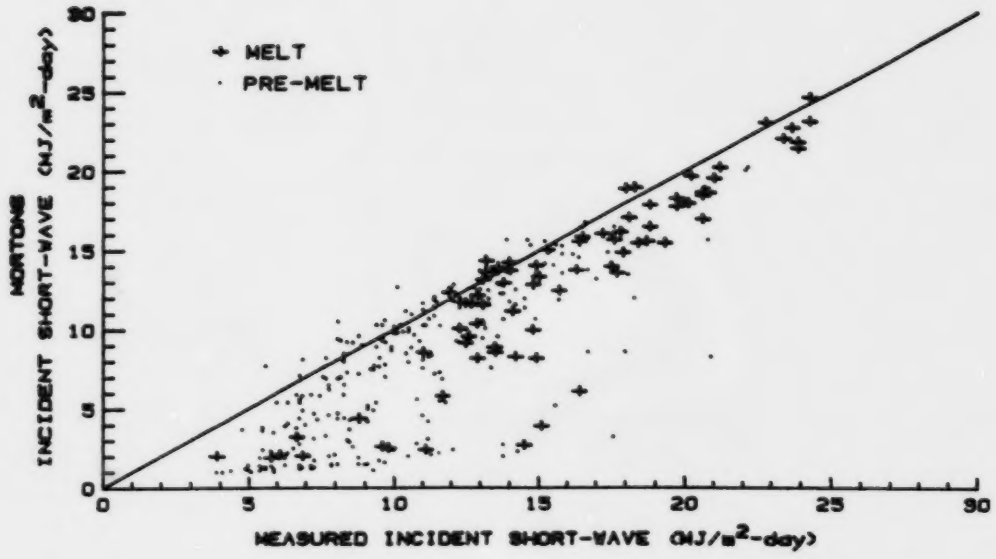

Figure 4.1 CRAE versus measured daily incident short-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78.

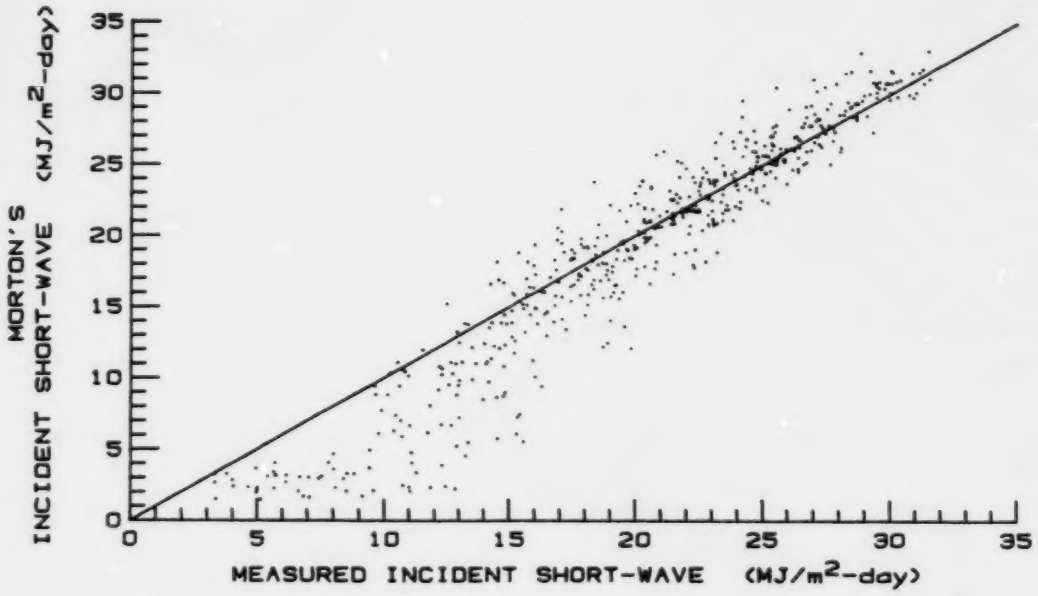

Figure 4.2 CRAE versus measured daily incident short-wave radiation fluxes at Bad Lake for Postmelt and Summer periods, 1981-83.

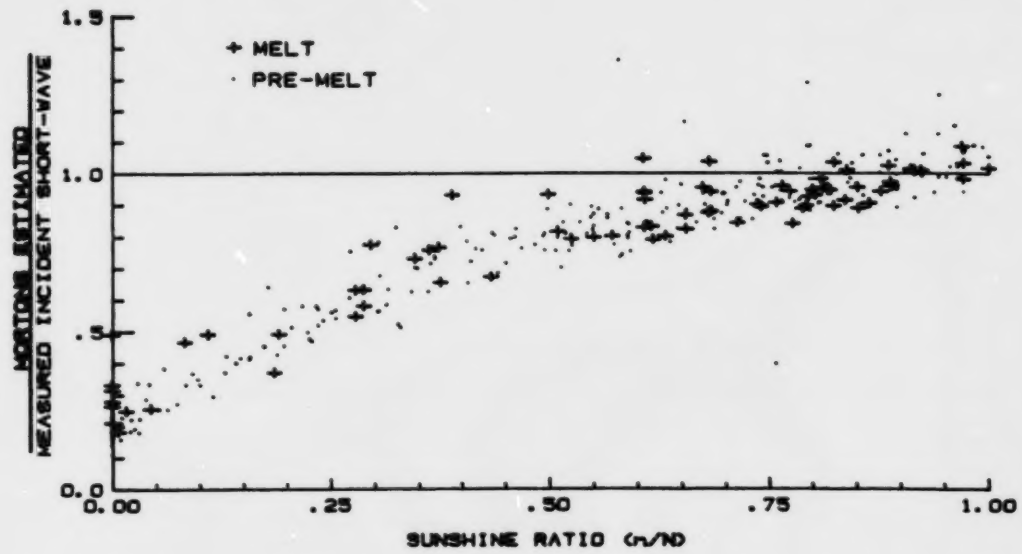

Figure 4.3 Ratio of CRAE to measured daily incident short-wave radiation against sunshine ratio for periods with snow cover, Bad Lake, 1972-78.

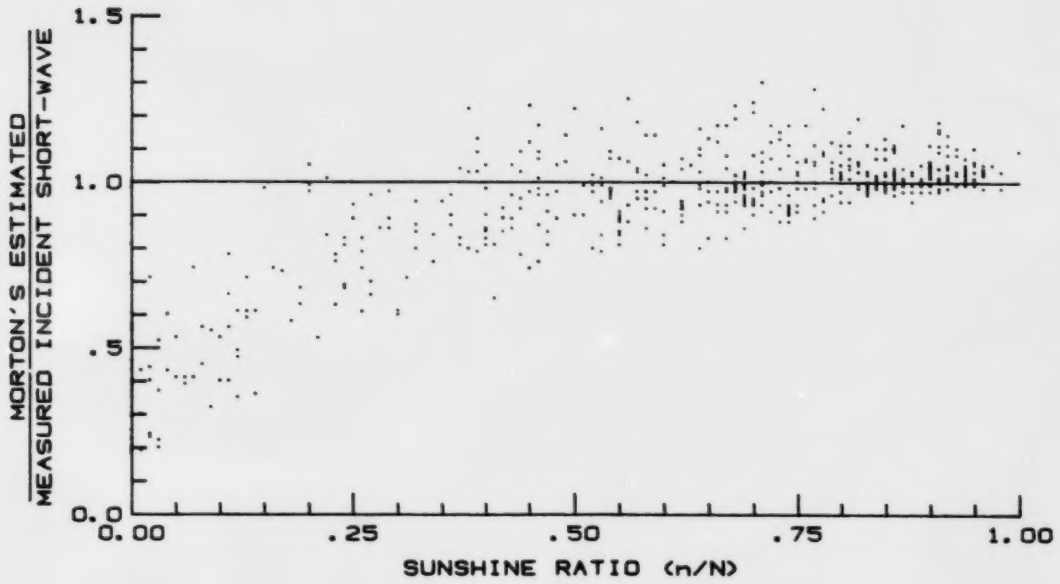

Figure 4.4 Ratio of CRAE to measured daily incident short-wave radiation against sunshine ratio for Postmelt and Summer periods, Bad Lake, 1981-83.

is derived. First the "zenith value of the dry season snow-free clear sky albedo" (A_{zd}) is calculated as a constant value for the station. A_{zd} varies from 0.11 to 0.17 and is a function of the mean annual precipitation, station elevation and latitude. This parameter is the only correction in the model for regional vegetation, approaching 0.17 for arid (and barren) climates and 0.11 for humid (and lushly-vegetated) climates. The calculated value of A_{zd} for Bad Lake is 0.13 indicating relatively lush vegetation, while for Saskatoon it is 0.11, the maximum on the side of lushness.

To arrive at the actual surface albedo, corrections are made to account for the possible effects of a wet season in an arid region, for the presence of snowcover, for the noon zenith angle of the sun, and for cloud cover. The daily albedo calculated by Morton's formulations is plotted against the albedo calculated from measured reflected and incident short-wave radiative fluxes at Bad Lake for the snow season in Fig. 4.5 and for the snow-free period in Fig. 4.6. Note, the grouping of points in Fig. 4.6 represents 486 data pairs. While a large number of albedos monitored during Premelt fall between 0.7 and 0.9, the model estimates range from 0.25 to 0.8. During Melt, measured albedos may decrease from 0.8 to 0.2, while Morton's estimates are usually at approximately 0.25. One can only conclude that Morton's calculations are inadequate for predicting snow surface albedos before and during snowmelt. For the snow-free conditions the range of albedo estimates is much smaller, from a definite minimum of 0.19 to approximately 0.28, while the observed values at Bad Lake fall between 0.13 and 0.23. Independent measurements of albedo near Saskatoon for various landuses during 1988 confirmed the range of observed albedo values in Fig. 4.6. The albedo of fallow ranged from 0.10 in a moist condition to 0.14 in the dry condition; albedos for a growing wheat crop ranged from 0.14 to 0.19 and showed only a minimal variation ($\mp$0.01) with sun angle during the day; and the albedo of a native grass was 0.18. All data suggest that the estimate of snow-free surface albedo given by the model is too high. The limited range and scatter in "modelled" albedos (Fig. 4.6) also suggest that use of a constant albedo may be equally valid as the estimate obtained with the model.

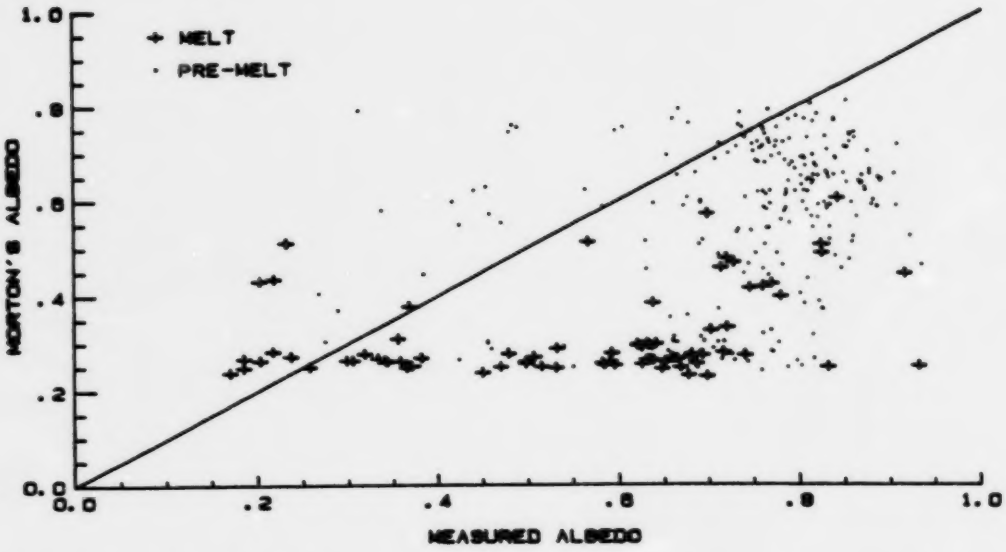

Figure 4.5 CRAE versus measured daily albedos at Bad Lake during periods with snow cover, 1972-78.

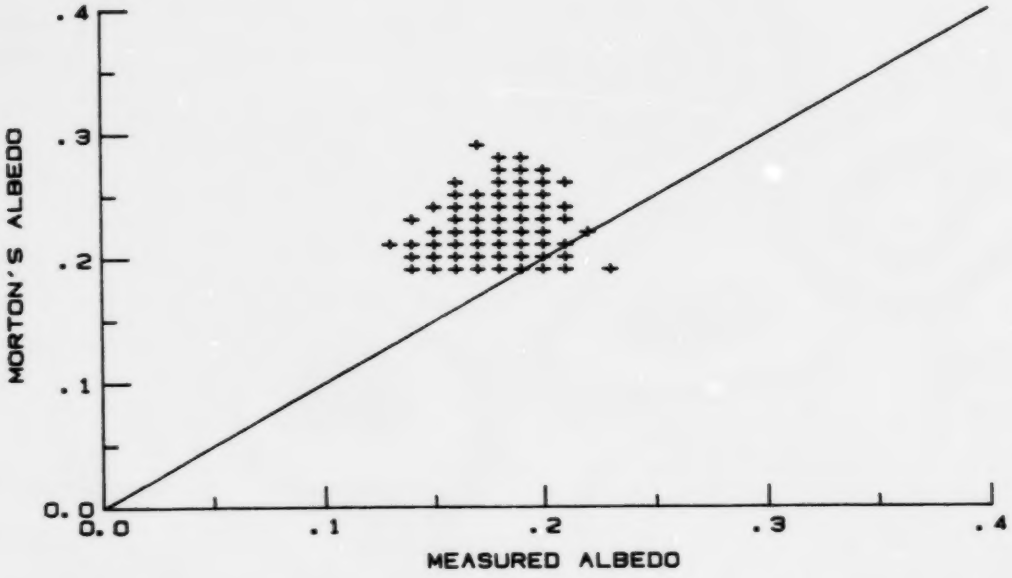

Figure 4.6 CRAE versus measured daily albedos at Bad Lake during Postmelt and Summer periods, 1981-83.

Daily estimates of net short-wave radiation (Q_{sn}) by the model for snow-free periods are compared to observed values in Fig. 4.7. The data show relatively large scatter and suggest that Morton's procedure generally results in an underestimate of this term and that the error is greatest for small values of Q_{sn}. This error is the result of an underestimate of the transmittancy of clouds and an overestimate of the albedo. The mean difference between the calculated and observed Q_{sn}-values is 1.41 MJ/m^2-d with a standard deviation of 1.96 MJ/m^2-d.

4.1.2 Long-wave radiation.

Net long-wave radiation is calculated by Morton's model as a single parameter indicating radiative loss and it is not possible to separately evaluate incoming and outgoing fluxes. The expression for net long-wave radiation is:

$$-Q_{ln} = \epsilon\sigma(T+273)^4[1-(0.71 + 0.007v_D p/p_s)(1 + \rho)] \qquad 4.4[C\text{-}26]$$

where ϵ is the surface emissivity, σ is the Stefan-Boltzmann constant, T is the mean air temperature, p and p_s are the mean station and sea-level atmospheric pressures, respectively, v_D is the atmospheric vapor pressure and ρ is the "proportional increase in atmospheric radiation due to clouds". It appears that the increase referred to is the downward long-wave flux originating from emission and reflection in the atmosphere. The factor ρ is primarily a function of the sunshine ratio and the vapor pressure. However, long-wave radiation emitted during night-time hours can be important and the sunshine ratio is not an ideal index of cloudiness for this period. Unfortunately, measurements of night-time cloudiness are not regularly observed at meteorological stations.

Morton uses a constant value of 0.92 for the surface emissivity. This is adequate for vegetated surfaces, but not for snow-covered surfaces. An emissivity in the range 0.97 to 1.0 would be more realistic.

The term $(0.71 + 0.007v_D p/p_s)$ in Eq. 4.4 was determined using data reported by Sellers (1965) by regressing 15 values for the ratio of clear-sky to black-body radiation against the ambient water vapor pressure. Refinement of the relationship may be possible using data recently published by Aase and

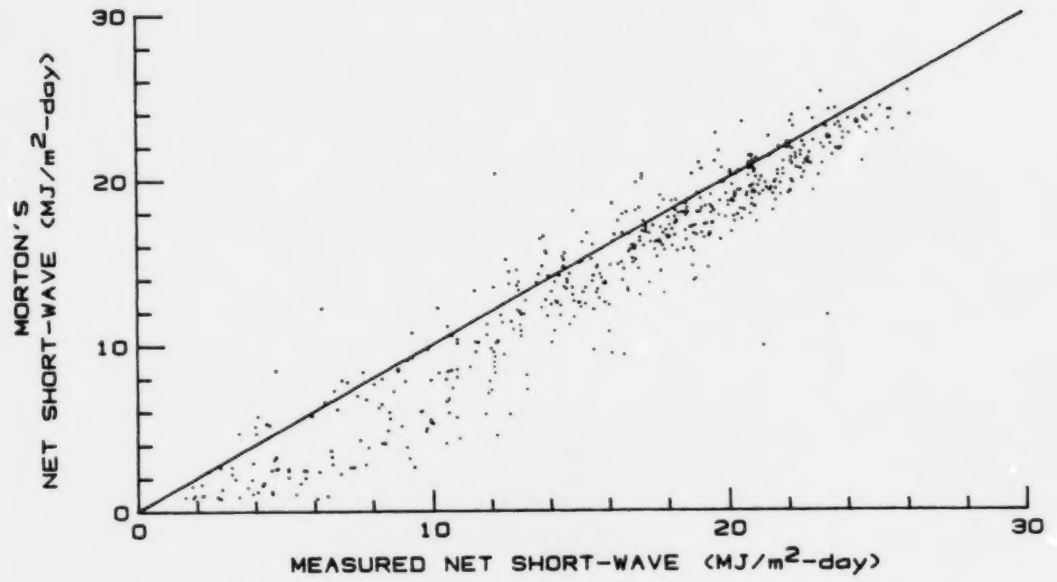

Figure 4.7 CRAE versus measured daily net short-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83.

Idso (1978), Brutsaert (1975) and Satterlund (1979).

Daily net long-wave radiation losses calculated by Eq. 4.4 are compared to measured daily net long-wave radiation, calculated as net all-wave less net short-wave, at Bad Lake for snow-covered periods (Fig.4.8) and for snow-free periods (Fig. 4.9). The data suggest that the model estimate of Q_{ln} is limited to a range of approximately 2 to 8 MJ/m^2-d; the measured values range from 0 to 14.3 MJ/m^2-d. The mean differences between the calculated and measured fluxes are: 0.85 MJ/m^2-d for Premelt, 0.54 MJ/m^2-d for Melt, and 1.80 MJ/m^2-d for the snow-free periods, with standard deviations of 1.25, 1.75, and 1.70 MJ/m^2-d, respectively. These large variances put in doubt the acceptability of model estimates for short periods.

4.1.3 Net all-wave radiation

The net all-wave radiation (Q_N) is calculated as the sum of the net short-wave and net long-wave fluxes (Eq. C-27). Daily values of Q_N calculated by the model are plotted against net all-wave radiation measured at Bad Lake in Figs. 4.10 and 4.11 for the snow-covered and snow-free periods, respectively. The results indicate that errors in individual components of the radiation budget do not compensate for each other in the net all-wave flux. For the snow-free periods, Fig. 4.11 translates both the effects of improper accounting for the transmittancy of clouds and albedo, as well as the limited range of the estimated long-wave loss. The data show the tendency for Morton's procedure to underestimate Q_N at low fluxes and to overestimate the value at large fluxes. Given the direct effect of Q_N on the estimate of evaporation, this tendency could translate into a seasonal trend for underestimatng evaporation during the early part of the snow-free period and overestimating evaporation during the summer months. The mean differences between the calculated and measured daily net all-wave fluxes are 0.22 MJ/m^2-d for Premelt, 1.65 MJ/m^2-d for Melt, and 0.39 MJ/m^2-d for snow-free conditions, with corresponding standard deviations of 2.65, 3.37, and 2.58 MJ/m^2-d, respectively.

The analysis was repeated for 5 and 10-day periods using data for the snow-free season and the standard deviations of the differences between the CRAE and measured net all-wave radiation for the two periods were 1.80 and 1.5 MJ/m^2-d, respectively. These findings demonstrate a decrease in the variance of the difference with increasing computational interval.

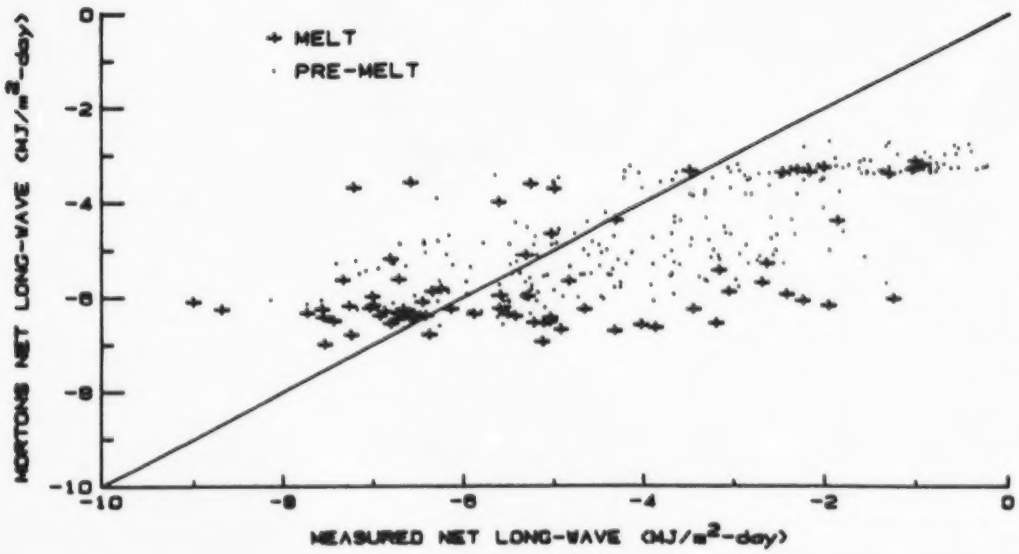

Figure 4.8 CRAE versus measured daily net long-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78.

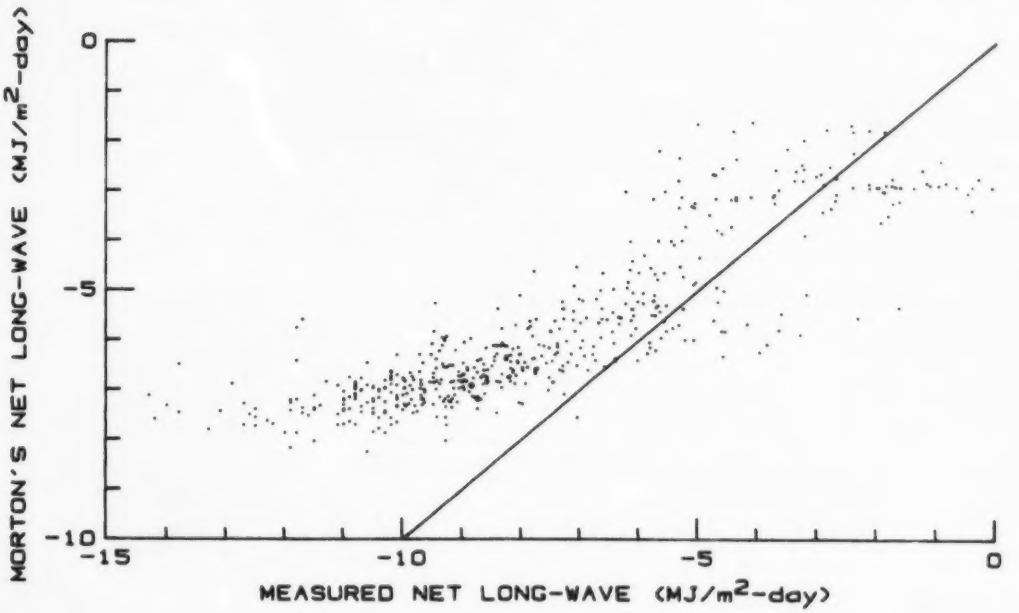

Figure 4.9 CRAE versus measured daily net long-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83.

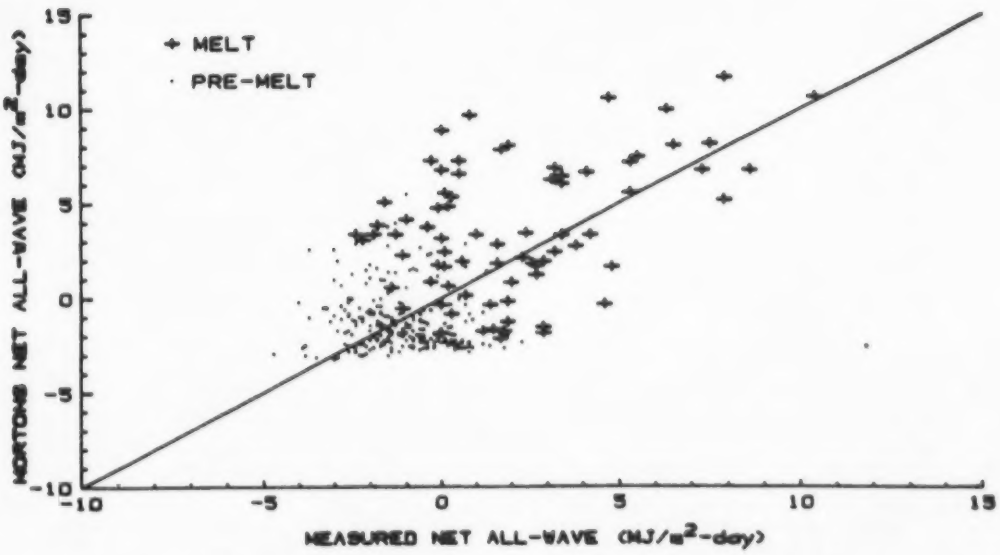

Figure 4.10 CRAE versus measured daily net all-wave radiation fluxes at Bad Lake for periods with snow cover, 1972-78.

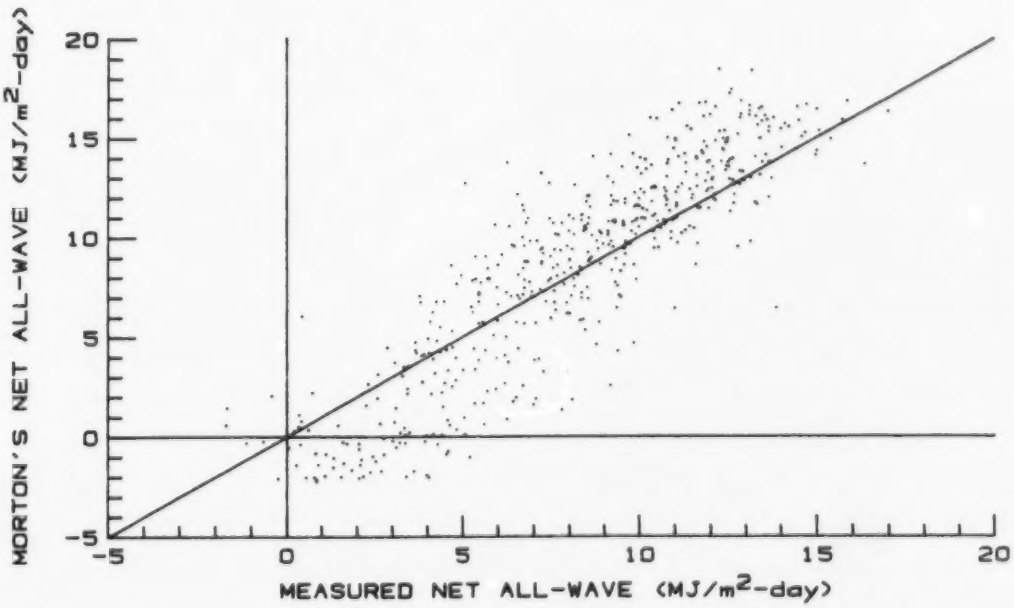

Figure 4.11 CRAE versus measured daily net all-wave radiation fluxes at Bad Lake for the Postmelt and Summer periods, 1981-83.

4.1.4 Summary.

The material in Sections 4.1.1 to 4.1.3 focussed on an assessment of the routines described by Morton (1983) for daily estimates of the radiative energy terms over snow-covered and snow-free surfaces. On the basis of the results of these analyses it is recommended that Morton's radiation algorithm not be used for estimating net radiation over daily time periods. In assessment of the procedures proposed by Morton it must be appreciated that most of the empirical relationships were derived from monthly data and the algorithms were developed during the period of time when microcomputers were not readily accessible. Any model which is structured around monthly data, uses relatively-simple algorithms for process simulation and is designed to give parameter estimates over an integrated spatial and temporal scale, should not be expected to give good estimates over short time intervals under a site-specific set of physiographical and climatic conditions.

4.2 Evaluation of the Vapor Transfer Coefficient.

In his development of CRAE, Morton assumed the vapor transfer coefficient to be independent of wind speed based on the following:

(1) The transfer coefficient increases with increases in both surface roughness and wind speed and wind speeds tend to be lower in rough areas than in smooth areas,

(2) The transfer coefficient increases with the instability of the atmosphere and its effect is more pronounced at low wind speeds than at high wind speeds, and

(3) Climatological observations of wind speed can lead to significant error because of local variations in exposure and instrument height.

With regard to these justifications, item 3 is a general statement which applies equally well to any climatological observation, be it wind speed, temperature or humidity. In order for the assumption of a transfer coefficient independent of wind speed to be valid, the offsetting effects described as items 1 and 2 must be of the same order of magnitude. That is: (a) an increase in magnitude of the transfer coefficient resulting from an increase in surface

roughness must be comparable to it's decrease in value resulting from the corresponding decrease in wind speed over the rougher terrain, and (b) the effect of changes in atmospheric stability must be comparable to the effect of changes in wind speed which accompany the changes in stability.

The vapor transfer equation, in its simplest form, is written as:

$$E = f_T(v_S - v_D), \tag{4.5}$$

where f_T is the vapor transfer coefficient, and v_S and v_D are the vapor pressures at the surface and instrument height, respectively. Using the aerodynamic formula, and assuming that the wind speed measurement is taken at a height of 10 m and the instrument height is 2 m, the vapor transfer coefficient can be written as:

$$f_T = 0.622\zeta k^2 \rho u / P \ln[(10 - d_o)/z_o] \ln[(2 - d_o)/z_o] \tag{4.6}$$

where ζ is the ratio of transfer coefficients for vapor and momentum, k is the von Karman constant, ρ is the air density, P is the atmospheric pressure, u is the wind speed and d_o and z_o are the displacement and roughness heights, respectively. ζ is the ratio of transfer coefficients for vapor and momentum, and is a function of atmospheric stability. Dyer and Hicks (1970) provide the following relationship:

$$\zeta = (1 - 16z/L)^{0.25}, \tag{4.7}$$

where z is the height and L is a scaling length introduced by Obukhov (1946), which is given by the expression:

$$L = u_*^3 C_p \rho T / kgH , \tag{4.8}$$

where u_* is the friction velocity, C_p is the heat capacity of the air, g is the acceleration due to gravity, and H is the sensible heat transfer rate.

d_o and z_o characterize the surface roughness and, according to Brutsaert (1982), can be expressed in in terms of the mean height of the vegetation, h,

as $d_0 = .67h$ and $z_0 = h/7.35$.

Using the above relationships the effects of wind speed, surface roughness and atmospheric stability on the vapor transfer coefficient can be examined. Figure 4.12 shows the vapor transfer coefficient plotted against wind speed for several conditions of surface roughness. In deriving these plots a mean sensible heat flux rate of 70 W/m^2 was assumed, and the effects of atmospheric stability on ζ were included. Also shown in the figure is the line representing the vapor transfer coefficient for the evaporation pan at Bad Lake using data for the 11-year period, 1973 to 1983 in Eq. 4.5. The data show that the effect of wind speed on the vapor transfer coefficient is least for very smooth surfaces; however, even for a short grass surface the transfer coefficient increases by a factor of 2.25 as the wind speed increases from 1 to 8 m/s. The effect of wind speed on f_T is even more pronounced for rougher surfaces; for a surface covered with 1-m vegetation (ie. a mature wheat crop) the transfer coefficient increases by a factor of 3.2 as the wind speed changes from 1 to 8 m/s. These calculations suggest that the assumption that the vapor transfer coefficient is independent of the wind speed is weak.

Figure 4.12 also shows that the effects of the surface roughness on the transfer coefficient are significant; for example, for a mean wind speed of 5 m/s, the transfer coefficient increases approximately by a factor of 5 as air passes from a short grass to a mature cereal crop.

Morton derived his vapor transfer coefficient by trial and error in conjunction with calibration of the expression used to calculate the wet-environment areal evapotranspiration. His analysis was based on monthly values for a number of sites, and the results (Fig. 10, Morton, 1983) show relatively little scatter. The fact that monthly periods (rather than shorter periods) are used contributes significantly to reducing scatter. Mean monthly wind speeds tend to show relatively little variation compared with daily values. This is demonstrated in observations at the Saskatchewan Research Council Meteorological site near Saskatoon. Table 4.1 shows the standard deviation of the 10-m wind speed and the ratio of the standard deviation of the mean to the mean wind speed for periods of various lengths. For daily values the standard deviation of the mean wind speed is equal to 53% of the mean daily wind, for monthly periods it is equal to 10% of the mean monthly wind, and for periods of

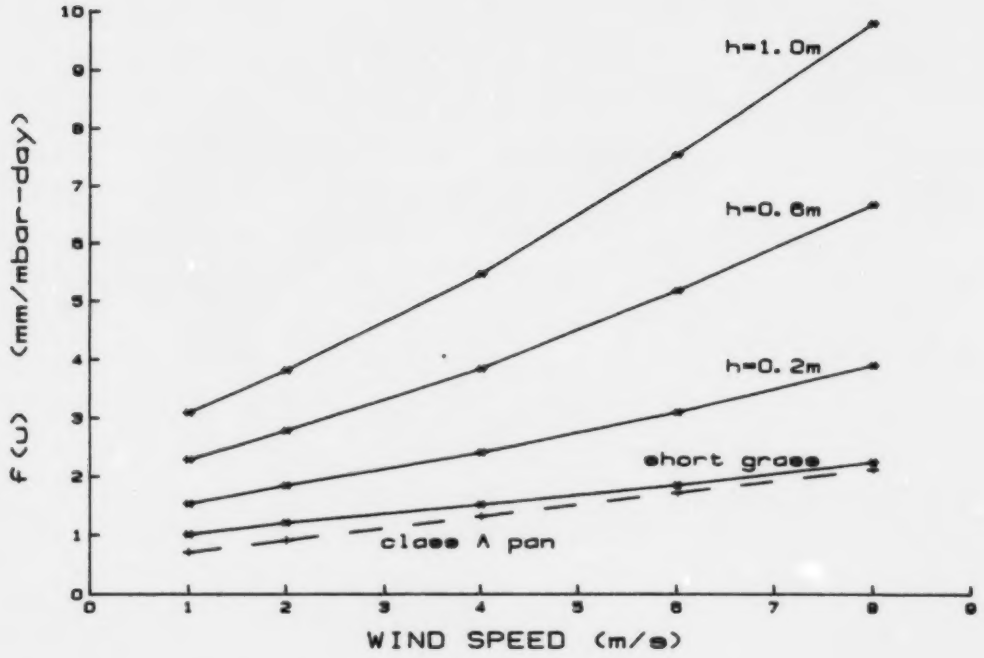

Figure 4.12 Vapor transfer coefficient plotted with wind speed for various vegetated surfaces and for a Class A evaporation pan.

90 days and longer it is approximately equal to 5% of the mean. Use of monthly periods significantly reduces the range of mean wind speeds and consequently the range of vapor transfer coefficients encountered with shorter time periods of observation.

Table 4.1 Standard deviation of the 10-m mean wind speed as a function of measurement period for Saskatoon (SRC) (1978 - 1987).

Period	Standard Deviation of mean wind speed m/s	% of mean wind speed
daily	2.12	53
5-day	1.56	39
10-day	1.25	31
monthly	0.40	10
3-month	0.24	6
6-month	0.20	5
yearly	0.20	5

One effect of assuming the vapor transfer coefficient is independent of wind speed is to limit the application of the model to periods of the order of one month or longer. However, monthly mean wind speeds can exhibit both annual and seasonal variations. At Saskatoon for example March, April and May (mean monthly wind speed of 4.5 m/s) tend to be more windy than July and August (mean wind speed = 3.5 m/s). Thus, if the vapor coefficient is calibrated for spring, the CRAE model will tend to overestimate evapotranspiration during summer.

Morton's observation that his results (calibration of the vapor transfer coefficient and the wet-environment areal evaporation) show less scatter for open-water surfaces than vegetative and land surfaces also confirms that surface roughness plays a significant role in the magnitude of the vapor transfer coefficient. The variability in the vapor transfer coefficient over smooth water surfaces can be expected to be less than that over rough, vegetated and land surfaces.

5. SENSITIVITY ANALYSIS

The CRAE model requires two levels of input: namely, one-time parameters describing the site of interest, and periodical meteorological data. The site specific parameters include the latitude of the station, the average atmospheric pressure (or the altitude of the station), and the average annual precipitation. The sensitivity of the model to these inputs will not be evaluated as it may be assumed that the user will (or should) have the correct values at his disposal.

The required meteorological data are temperature, humidity, and insolation. The model allows for several options as to the type of humidity and insolation data it will accept. Humidity data may be input as dewpoint temperature, vapor pressure, or relative humidity and insolation data may be input as the ratio of observed to maximum possible sunshine duration, the observed sunshine duration, or the observed global radiation. Whichever option is used, the model performs the necessary conversions to make the data compatible with the working equations. In the following assessment, the sensitivity of the model performance to variations in the input of air temperature, vapor pressure, and sunshine duration is evaluated. The data set used for this assessment are daily mean values for the period March to September, 1981 at Saskatoon. A five-day averaging period is used.

5.1 Temperature.

Air temperature input is used in the calculation of several components of the model. It first appears in the calculation of the net long-wave radiation loss (Eq. C-26, Morton, 1983), where a positive error in the temperature data causes an overestimate of this term, and hence, an underestimate of the net all-wave radiation. Net radiation and air temperature both appear in the calculation of the equilibrium temperature, T_p (Eqs. C-32, C-33), where their effects are offsetting: the lower net radiation lowers the estimate of T_p, while the higher temperature raises it. The net effect appears to be an underestimate of the equilibrium temperature, and hence of the temperature gradient, $T_p - T$. The net radiation and the temperature gradient are used to calculate the

potential evapotranspiration, E_{TP} (Eq. C-36) and again the effects of each term are somewhat offsetting, and the net result is an overestimate of the potential evapotranspiration. The lower equilibrium temperature, T_p, means a higher value for the net radiation at the equilibrium temperature, R_{TP}, which in turn results in an overestimate of the wet-environment areal evapotranspiration, E_{TW} (Eq. C-38).

In the complementary relationship (Eq. C-39), overestimates in both E_{TP} and E_{TW} have offsetting effects. The net overall effect of a positive error in the temperature data is an underestimate of the actual evapotranspiration.

To evaluate the sensitivity to temperature inputs, the CRAE model was run with a systematic error of $\mp 0.5^{o}C$ in the temperature data. (This error is approximately equivalent to the standard error of the estimate obtained when the daily mean temperature obtained from the mean of 24 hourly values is compared with the mean of the maximum and minimum temperatures.) Figure 5.1 shows that for the sample data used, a systematic error of $\mp 0.5^{o}C$ in temperature results in $\mp 6\%$ change in the estimate of seasonal evapotranspiration.

5.2 Humidity.

In the CRAE model, vapor pressure appears in the calculation of the net radiation (albedo, transmissivity, increase in the atmospheric radiation due to clouds) and the net long-wave radiation loss. The net result of a positive error in vapor pressure measurement is an overestimate of the net radiation. The increased net radiation and higher vapor pressure combine to provide an overestimate of the equilibrium temperature, T_p, and of the temperature gradient, $T_p - T$. However, the effects of an increase in net radiation and in temperature gradient tend to offset each other in the estimate of the potential evapotranspiration, E_{TP} (Eq. C-36) so as to cause an underestimate of this term. The higher equilibrium temperature results in an underestimate of the net radiation at the equilibrium temperature, R_{TP}, and consequently of the wet-environment evapotranspiration, E_{TW} (Eq. C-38).

Figure 5.1 Effect of a systematic error of ∓0.5°C in temperature on the CRAE estimate of evapotranspiration at Saskatoon, 1981.

In the complementary relationship (Eq. C-39), the effects of underestimates in both E_{TP} and E_{TW} tend to be offsetting. However, because the error in E_{TW} is larger, the net result of a positive error in vapor pressure is an overestimate of the actual evapotranspiration.

The most commonly-used methods of humidity measurement are the Lithium-chloride dewcell and the hygrothermograph and the accuracies generally associated with these instruments are ~3% and ~5% R.H., respectively. To test the sensitivity of the CRAE model to humidity inputs, the vapor pressure data were offset by an amount equivalent to a ∓5% error in relative humidity. This is equivalent to a ∓0.3 mb error at 0°C, and a ∓1.7 mb error at 20°C. Figure 5.2 shows that a systematic error of ∓5% in R.H. would lead to a ∓16% difference in the estimate of seasonal evapotranspiration.

5.3 Insolation and Radiation.

The insolation input appears only in the calculation of net radiation. The effect of a positive error in the insolation data is an overestimate of net radiation which results in an overestimate of both the potential, E_{TP}, and the wet-environment evapotranspiration, E_{TW}. The net result is that the actual evapotranspiration is overestimated.

Insolation data are recorded in hours of bright sunshine, to the nearest 0.1 h. The CRAE model was run with a systematic error of ∓0.2 h in the insolation data. Figure 5.3 shows that a systematic error of 0.2 h in the sunshine duration results in a 3% change in the seasonal estimate of actual evapotranspiration.

In Section 3.1 it is shown that the net all-wave radiation estimates were subject to relatively large errors. The standard deviations of the difference between calculated and measured net radiation for 5- and 10-day periods were 1.80 and 1.50 MJ/m^2-d, respectively. To test the effects of errors in net radiation, the CRAE model was run using uncorrected input data and a systematic error equivalent to ∓1.50 MJ/m^2-d was introduced once the net radiation was calculated. The result of this test, Fig. 5.4, shows that the systematic error in net radiation (∓1.50 MJ/m^2-d) changes the seasonal estimate of areal evapotranspiration by 24%. The test confirms the importance of the radiation

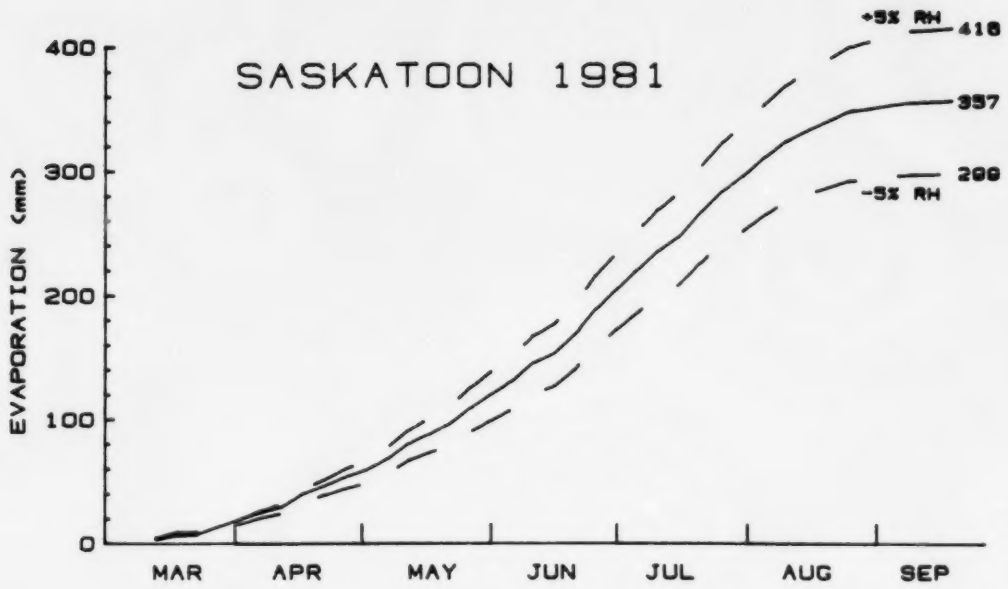

Figure 5.2 Effect of a systematic error of ∓5% in relative humidity on the CRAE estimate of evapotranspiration at Saskatoon, 1981.

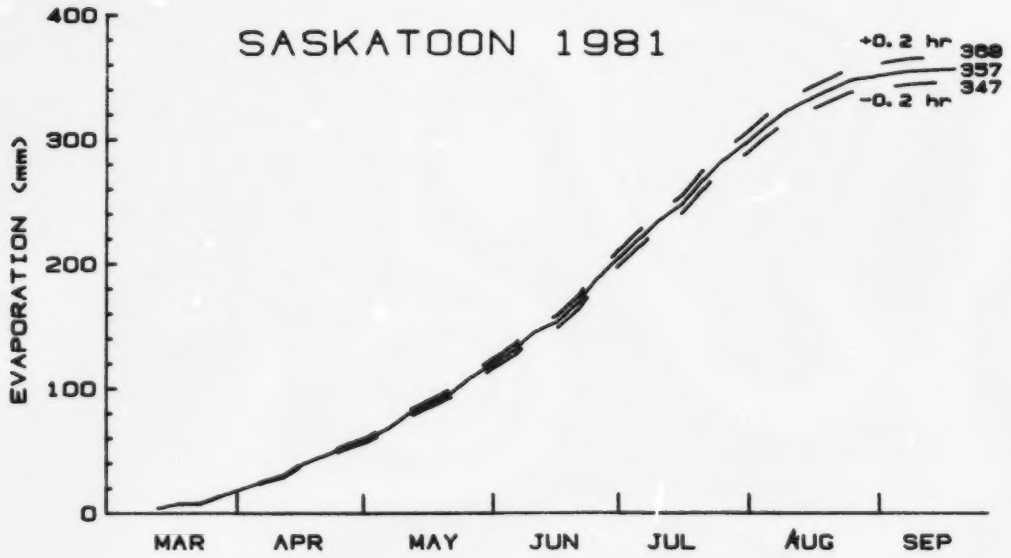

Figure 5.3 Effect of a systematic error of ∓0.20 h in sunshine duration on the CRAE estimate of evapotranspiration at Saskatoon, 1981.

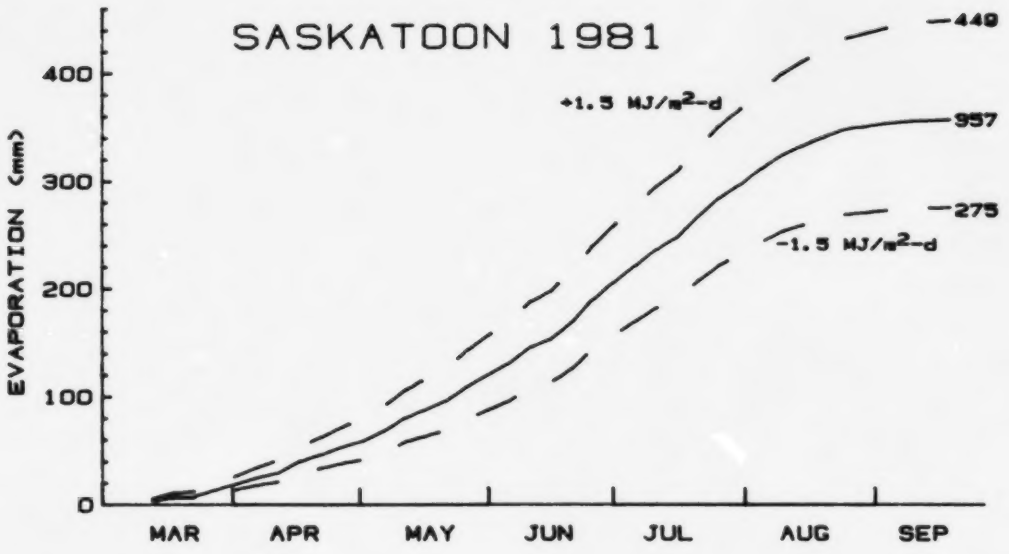

Figure 5.4 Effect of a systematic error of ∓1.5 MJ/m^2-d in calculated net all-wave radiation on the CRAE estimate of evapotranspiration at Saskatoon, 1981.

importance of the radiation flux to the estimate of evaporation, and the necessity to provide an accurate estimate of this term.

5.4 Averaging Period.

To study the effects of the time interval over which data are averaged (Averaging Period) on the estimates of evapotranspiration, the mean values of atmospheric variables calculated from daily means for 5-, 10- and 30-day periods were used as input to the model. Table 5.1 compares the 30-day evapotranspiration totals at Bad Lake in 1981 and 1983 with different averaging periods.

Table 5.1 Thirty-day evapotranspiration totals (mm) in 1981 and 1983 at Bad Lake, as calculated by the CRAE model using averaging periods of 5-, 10-, and 30-days.

Year	Period	Averaging Period (days)			
	Julian day	30	10	5	1
1981	63 - 92	18.9	17.7	20.2	17.6
	93 - 122	10.8	11.1	11.1	16.1
	123 - 152	18.6	25.2	28.2	30.5
	153 - 182	78.6	78.4	78.4	78.6
	183 - 212	82.7	83.6	81.5	80.2
	213 - 242	55.1	55.3	55.0	55.8
	Total	264.7	271.3	274.4	278.8
1983	70 - 99	28.3	15.0	18.7	18.4
	100 - 129	45.5	46.9	38.0	34.0
	130 - 159	61.9	61.5	61.5	59.4
	160 - 189	84.3	86.0	80.8	74.4
	190 - 219	93.4	93.3	92.2	88.3
	220 - 249	42.1	40.7	42.4	40.7
	Total	355.5	343.4	333.6	315.2

The data in the Table shows seasonal evapotranspiration increasing with a decrease in averaging period in 1981 whereas in 1983 the trend is reversed.

Closer inspection of the data reveals that in 1981 the largest differences between the 30-d totals for the averaging periods occurred in the Julian period 123 - 152 days, while in 1983 they occurred in the first two periods, 70 -129 days. For the majority of the 30-d intervals there is no discernible systematic variation in totals due to the choice of averaging period.

The Julian-day period 123 - 152 days in 1981, was marked by an abrupt change in dominant weather in the interval. The first half of the period was cool and dry, whereas the last part was warm and more humid. In 1983, in the first 60 days the temperature fell below freezing on several days and there were frequent periods of rain or snow. If an abrupt change in the weather occurs during an averaging period, the mean values of temperature and vapor pressure are less likely to produce a representative estimate of evapotranspiration because the model responds in a non-linear fashion to such changes. Conversely, if no large, abrupt changes occur, the choice of averaging period appears to have little effect on the estimate of evapotranspiration, and the correction procedure suggested by Morton, 1983 (Section C.2.5) for providing daily estimates is probably not needed.

6. COMPARISON OF MODEL ESTIMATES OF AREAL EVAPOTRANSPIRATION WITH ESTIMATES OBTAINED FROM MONITORED SOIL MOISTURE BALANCES.

Estimates of areal evapotranspiration by CRAE were compared with evapotranspiration estimates from soil water balances monitored at the Bad Lake watershed and at Saskatoon in the period 1981 to 1984. At each location, soil moisture data were available from representative sites in fallow, stubble and spring wheat and at Saskatoon measurements were also available at a site in native grass. The measurement period extended from the disappearance of the seasonal snowcover up to the time of harvest of the spring wheat crop.

Changes in soil moisture were calculated from measurements of changes in soil density made with a two-probe density meter to a depth of 1.4 m. The method and equipment have been described by Granger et al. (1984) and Gray and Granger (1986). Repeatability tests of the system in the field gave a standard error of estimate of 2.5 mm of moisture in a 1-m profile. The sum of the precipitation plus the soil moisture change observed in a measurement interval was taken as the estimate of evapotranspiration.

Figures 6.1 to 6.3 compare CRAE and "measured" estimates of evapotranspiration from wheat and fallow at Bad Lake in 1981, 1982 and 1983, and Figs. 6.4 to 6.7 show estimates for wheat, grass and fallow at Saskatoon in 1981, 1982, 1983 and 1984. Figures 6.8, 6.9 and 6.10 summarize these comparisons as plots of "modelled" versus "measured" evapotranspiration rates at corresponding times during Postmelt and Summer months. The CRAE estimates were generated using a 5-d averaging period and the "measured" estimates are the average rates for sites in the same landuse. From these results it can be observed:

(1) <u>In Postmelt</u>- before the seeding of annual crops (~mid-May):

  (i) the observed differences between evapotranspiration from fallow and stubble are not large (see Figs. 6.1-6.4).

  (ii) CRAE consistently overestimates actual evapotranspiration (eg. see Fig. 6.8). The overestimation is most likely due to the assumption in the model that the soil heat flux is negligible.

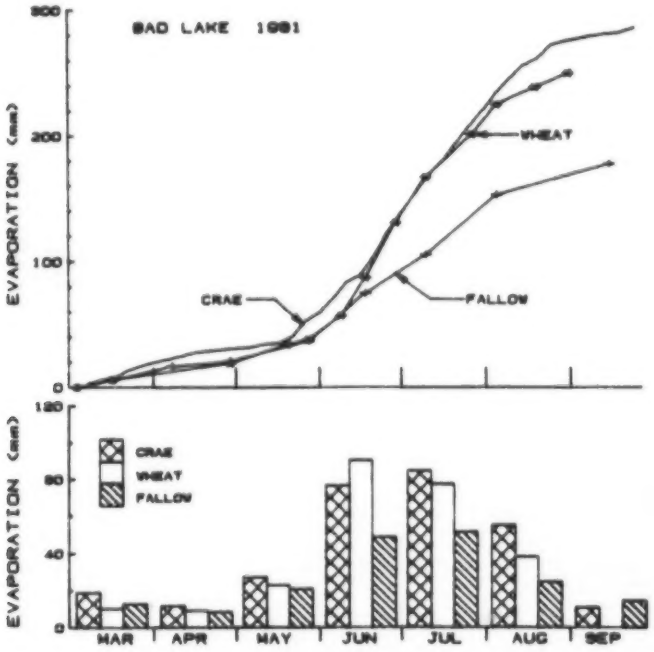

Figure 6.1 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1981.

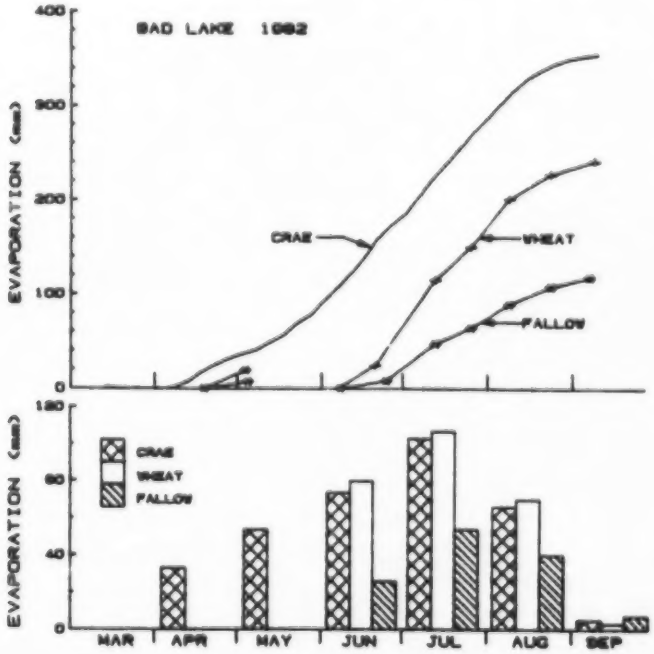

Figure 6.2 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1982.

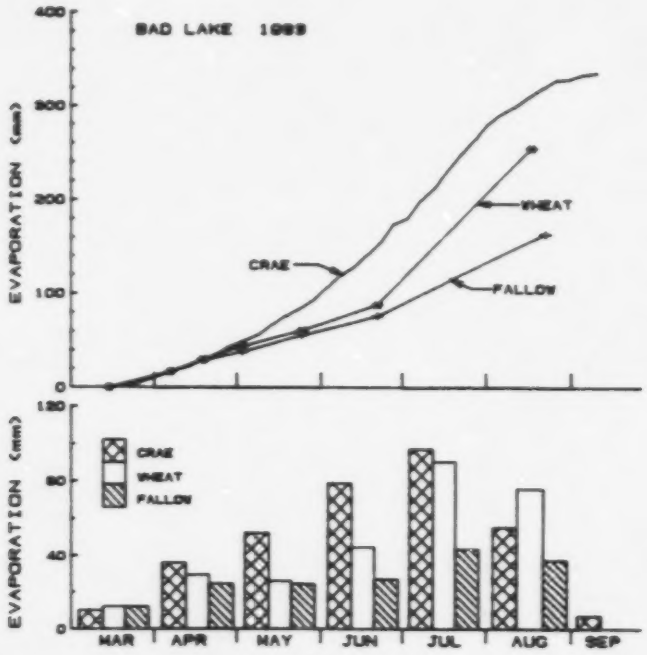

Figure 6.3 CRAE and average measured evapotranspiration from wheat and fallow at Bad Lake, 1983.

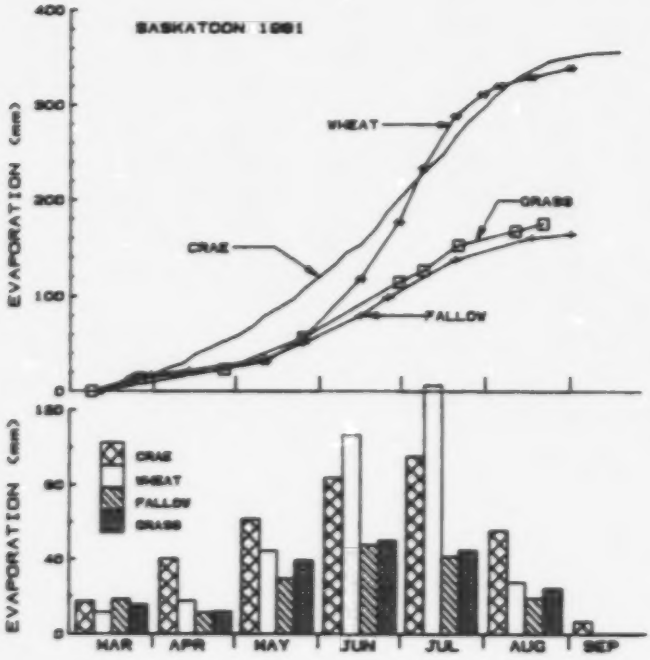

Figure 6.4 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1981.

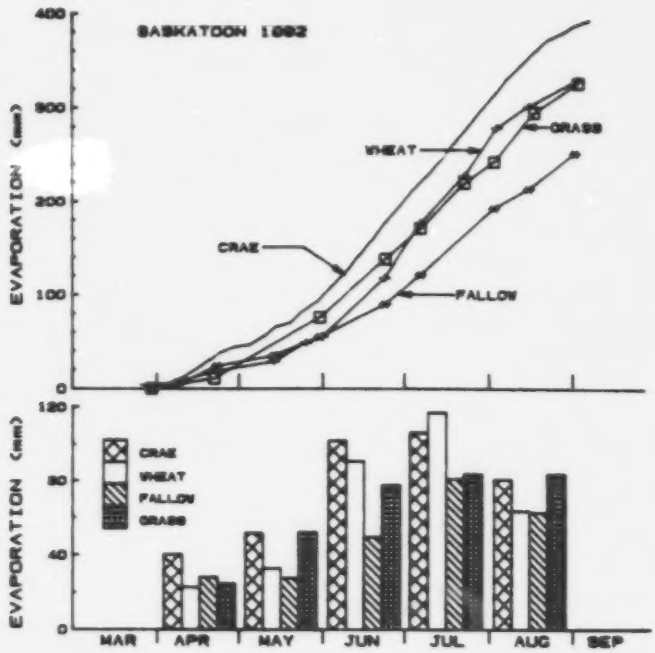

Figure 6.5 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1982.

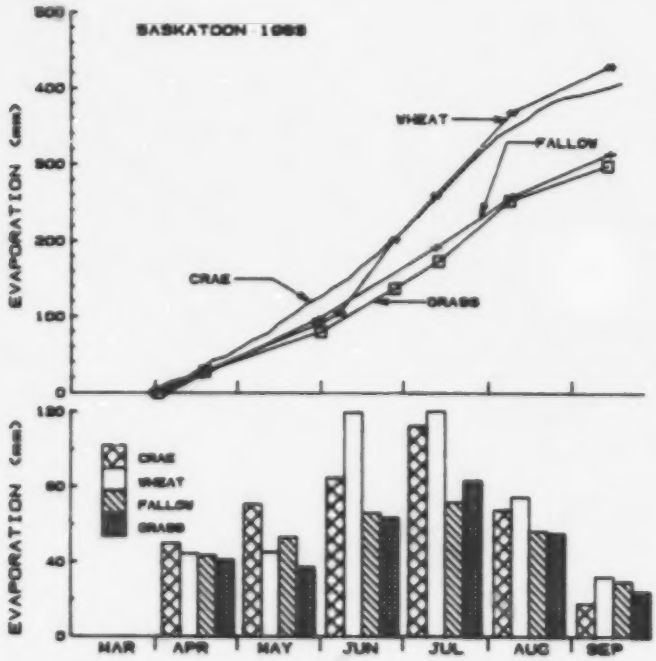

Figure 6.6 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1983.

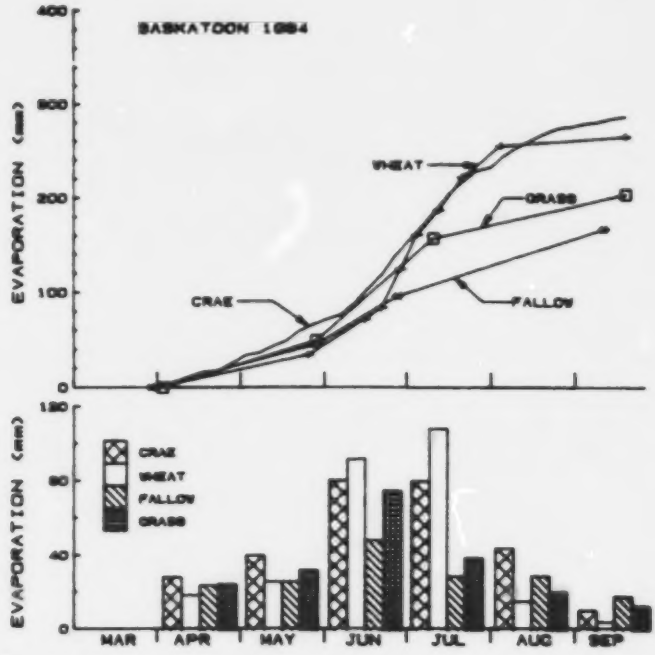

Figure 6.7 CRAE and average measured evapotranspiration from wheat, grass and fallow at Saskatoon, 1984.

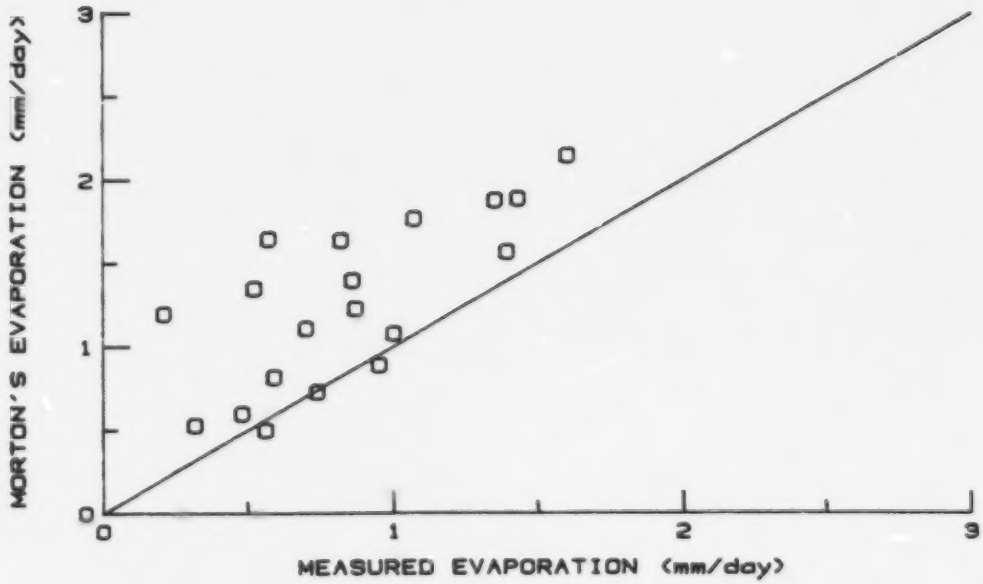

Figure 6.8 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration for March, April and May at Bad Lake (1981-83) and Saskatoon (1981-84).

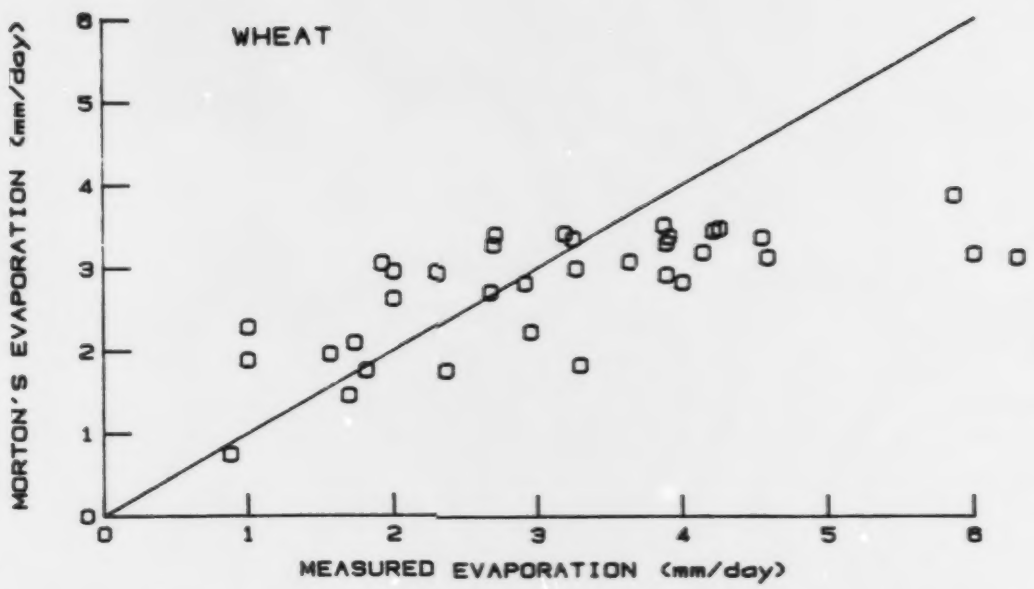

Figure 6.9 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration from wheat for June, July and August at Bad Lake (1981-83) and Saskatoon (1981-84).

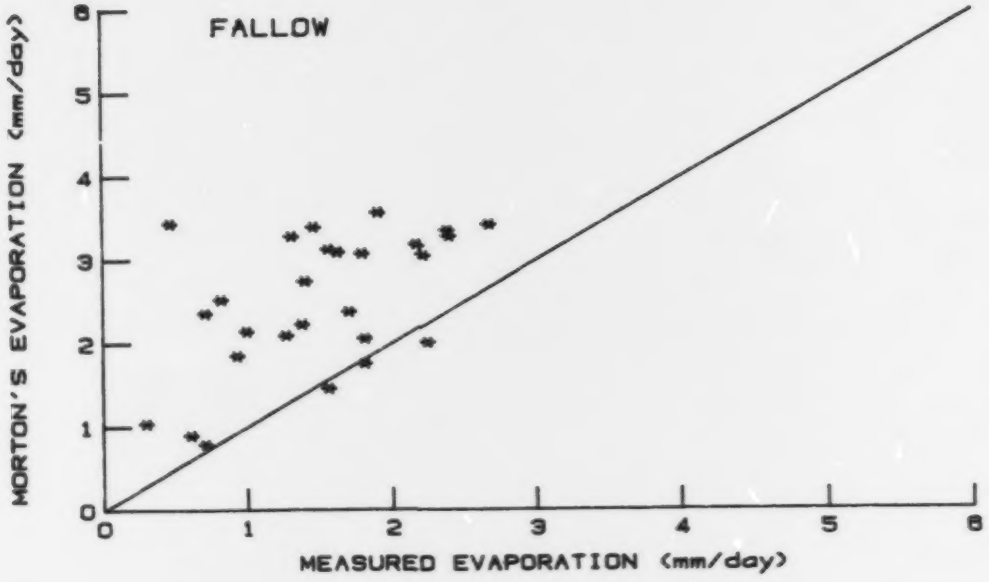

Figure 6.10 Estimation of evapotranspiration by CRAE plotted against measured evapotranspiration from fallow for June, July and August at Bad Lake (1981-83) and Saskatoon (1981-84).

While a soil is frozen, a significant portion of the available energy (net radiation) will be apportioned to soil to provide heat for the phase change and soil warming. The mean difference between "modelled" and "measured" evapotranspiration estimates during Postmelt is 0.42 mm/d and the standard deviation of the difference is 0.35 mm/d.

(2) <u>During the summer months</u>:

(i) Evapotranspiration from native grass follows more closely the pattern from fallow than from wheat (see Figs. 6.4 and 6.6).

(ii) CRAE underestimates evapotranspiration from wheat, particularly at high evapotranspiration rates (Fig. 6.9) and overestimates evapotranspiration from fallow (Fig. 6.10). The mean differences between CRAE and measured evapotranspiration were -0.38 mm/d on wheat and 1.02 mm/d on fallow with standard deviations of the difference of 1.04 and 0.74 mm/d respectively for the landuses.

These findings suggest that the CRAE model is not a good predictor of evapotranspiration from field-sized areas. To test its' performance for simulating evapotranspiration from larger areas the observed values were weighted to provide an estimate of "Regional" evapotranspiration. For this calculation it was assumed that a "region" has 55% of its' land area in wheat and 45% in fallow/grass. The comparisons of the CRAE and "Regional" estimates of evapotranspiration are presented in Figs. 6.11 through 6.17. These data exhibit a consistent trend for the model to overestimate "Regional" evapotranspiration during Postmelt and the cumulative effect of these monthly overestimations can produce an appreciable difference between the "modelled" and "measured" seasonal totals. During the summer a similar trend for the model to overestimate "Regional" evapotranspiration is evident in the data for Bad Lake (Figs. 6.11 - 6.13), whereas at Saskatoon (Figs. 6.14 - 6.17) there are months when the model overestimates and others when it underestimates the "Regional' value. Conversely, at each location there are months when the values agree favourably, eg. Bad Lake - June, 1981; Saskatoon - June and July 1981 (Fig. 6.14) and July, 1984 (Fig. 6.17).

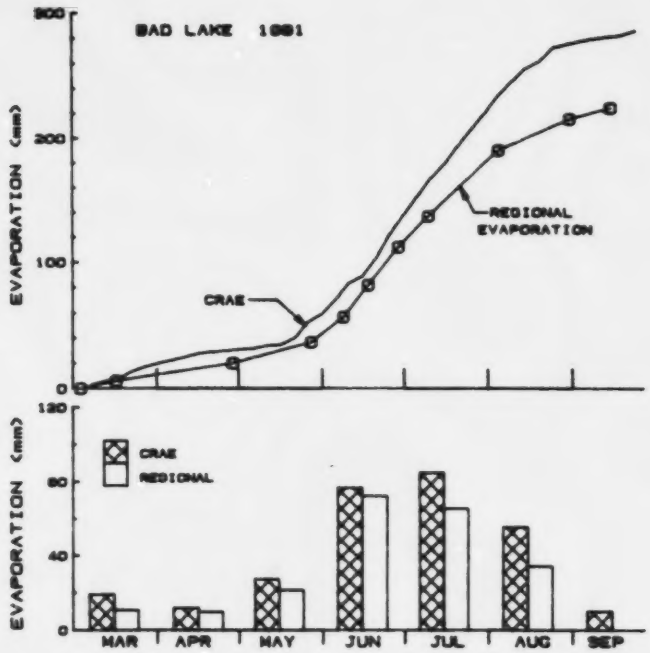

Figure 6.11 CRAE and measured regional evapotranspiration at Bad Lake, 1981.

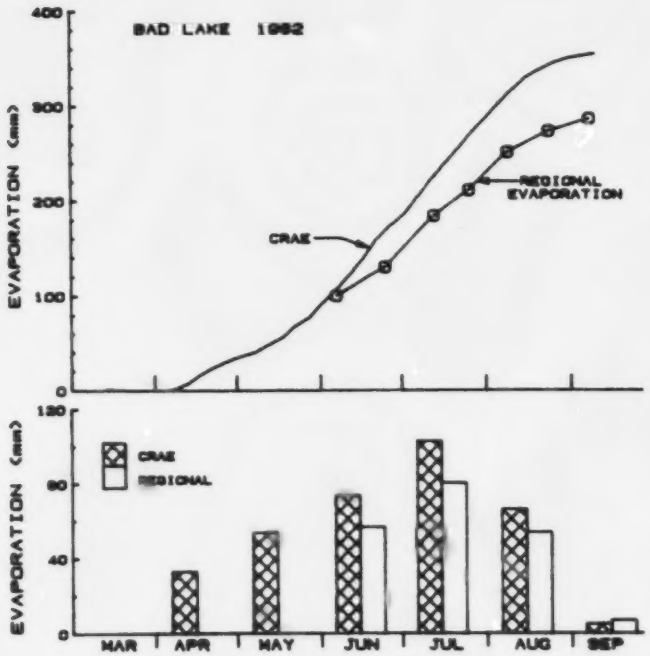

Figure 6.12 CRAE and measured regional evapotranspiration at Bad Lake, 1982.

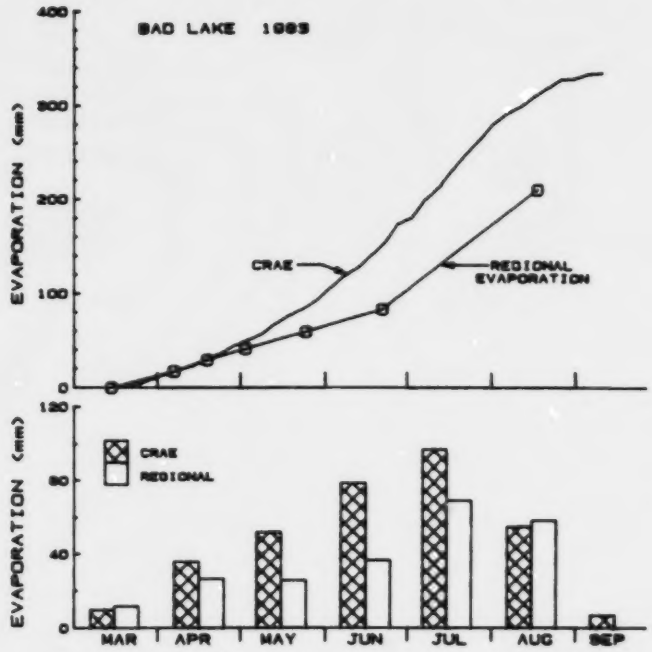

Figure 6.13 CRAE and measured regional evapotranspiration at Bad Lake, 1983.

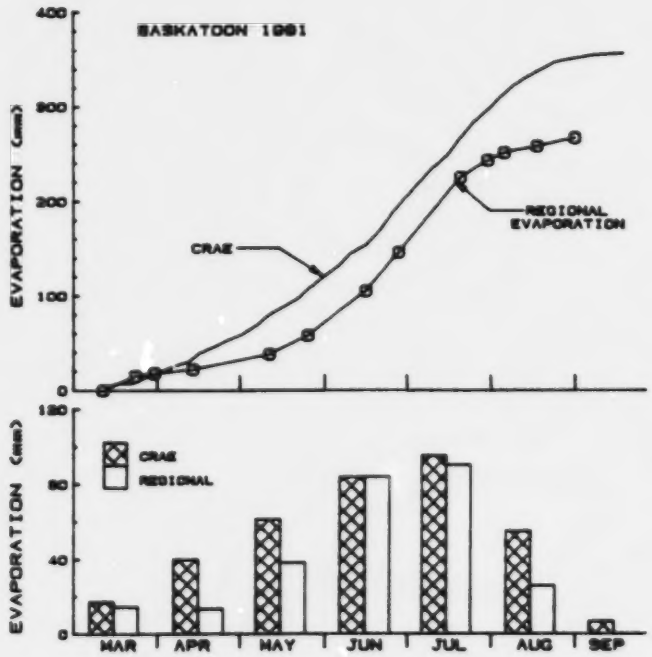

Figure 6.14 CRAE and measured regional evapotranspiration at Saskatoon, 1981.

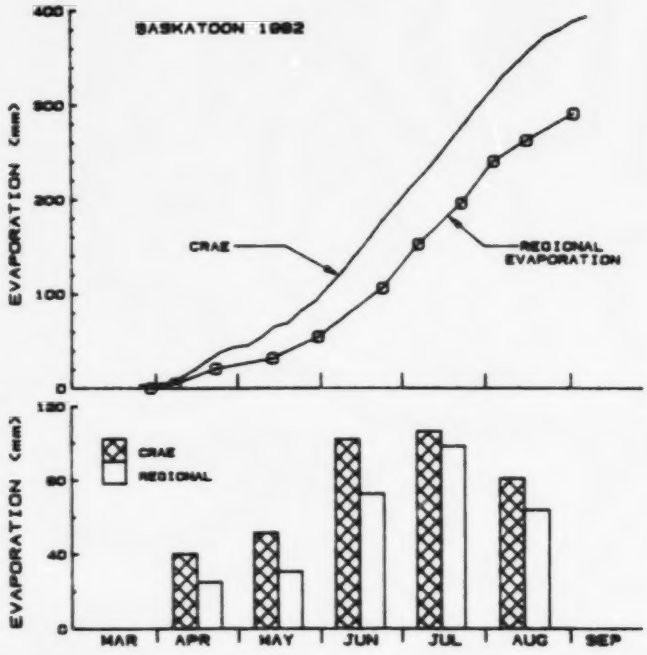

Figure 6.15 CRAE and measured regional evapotranspiration at Saskatoon, 1982.

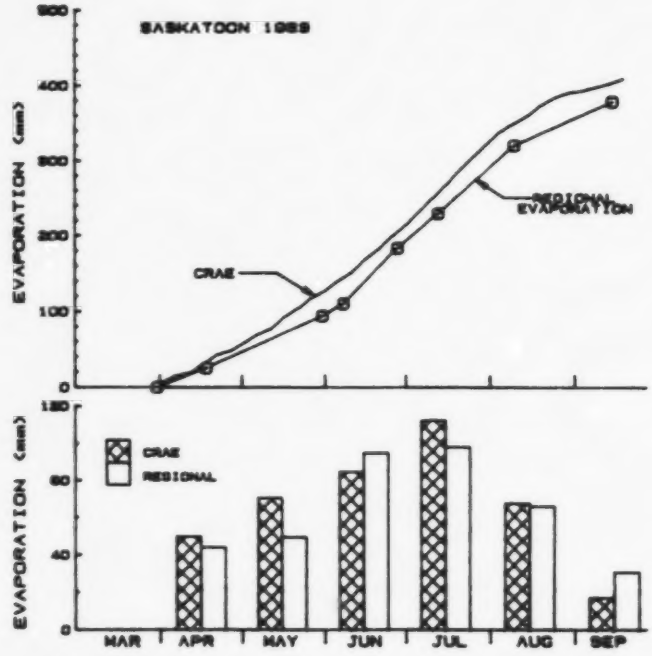

Figure 6.16 CRAE and measured regional evapotranspiration at Saskatoon, 1983.

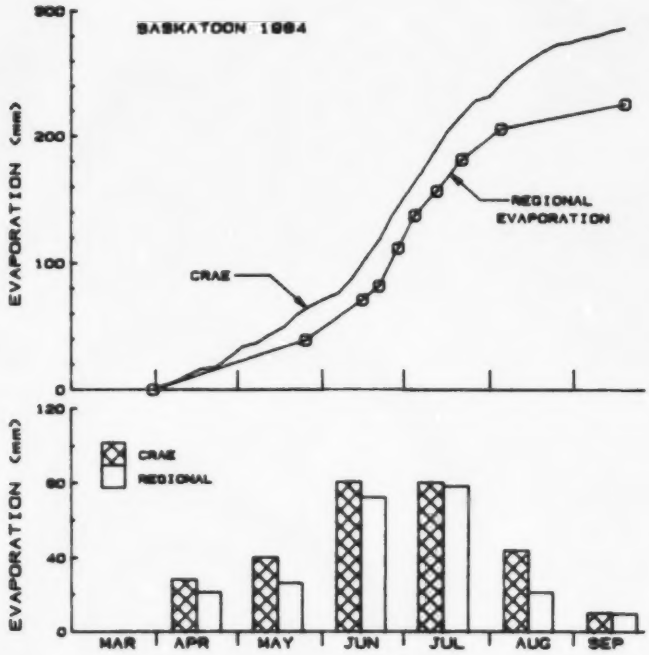

Figure 6.17 CRAE and measured regional evapotranspiration at Saskatoon, 1984.

In Section 4.2 it is suggested that Morton's assumption that the vapor transfer coefficient is independent of the wind speed is questionable. Closer examination of the wind regimes and evapotranspiration for June, July and August showed the model overestimated "Regional" evapotranspiration in months with light winds and underestimated the value in months with high winds. It was therefore anticipated that the "performance" of the model may be affected by it's inability to account for the effect of wind on the evaporation process. To determine whether closer agreement between "modelled" and "measured" estimates could be obtained by including the effects of windspeed and surface roughness on the vapor transfer coefficient, a correction for windspeed was incorporated into the system. The procedure used for this correction is described in Section 4.1 in conjunction with the derivation of Fig. 4.12. Figure 6.18 shows the results of the correction on the CRAE estimates of evapotranspiration for Bad Lake, 1981. In 1981, the mean wind speed at the station in June was slightly above average, while for July and August the mean

wind speeds were below the average. Fig. 6.18 shows that the adjustment significantly improves the model's simulation of the Regional evapotranspiration in July and August when the wind speeds were low.

Figures 6.19 and 6.20 show "Regional" estimates of the average evapotranspiration rate by CRAE for June, July and August at Bad Lake (1981, 1982) and Saskatoon (1981, 1982) plotted with measured rates. The plots differ in that the CRAE estimates in Fig. 6.20 are corrected for wind. Comparing these data it is obvious that the "wind correction" has improved the association between "modelled" and "measured" evapotranspiration estimates apppreciably. The mean differences between the parameters with and without the wind correction are 0.60 mm/d and 0.18 mm/d respectively and the corresponding standard deviations of the differences are 0.73 mm/d and 0.50 mm/d.

Even though the preceding comparisons use a 5-day averaging period, based on the results given in Table 5.1 it is anticipated that the incorporation into CRAE of a function which would account for the effects of wind on the vapor transfer coefficient in CRAE would also significantly improve its' simulation of monthly evapotranspiration.

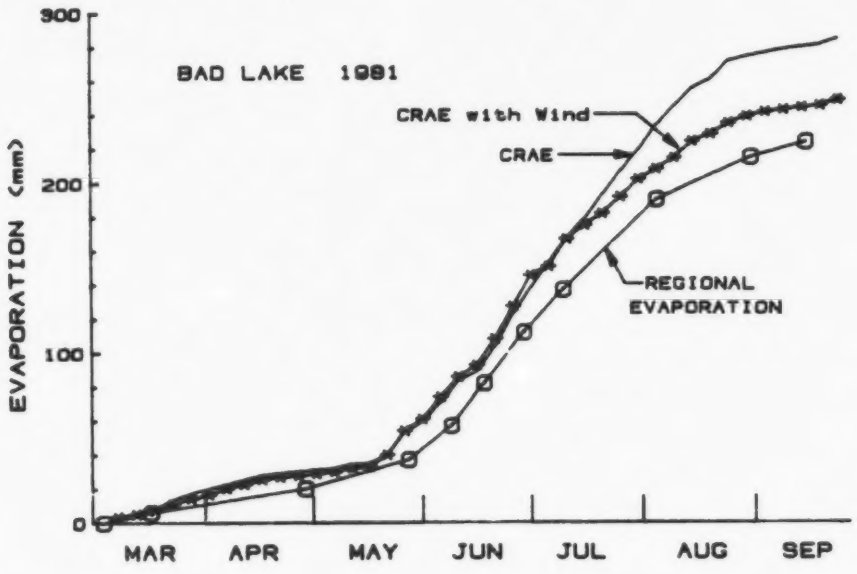

Figure 6.18 CRAE using a wind function for vapor transfer, and measured regional evapotranspiration at Bad Lake, 1981.

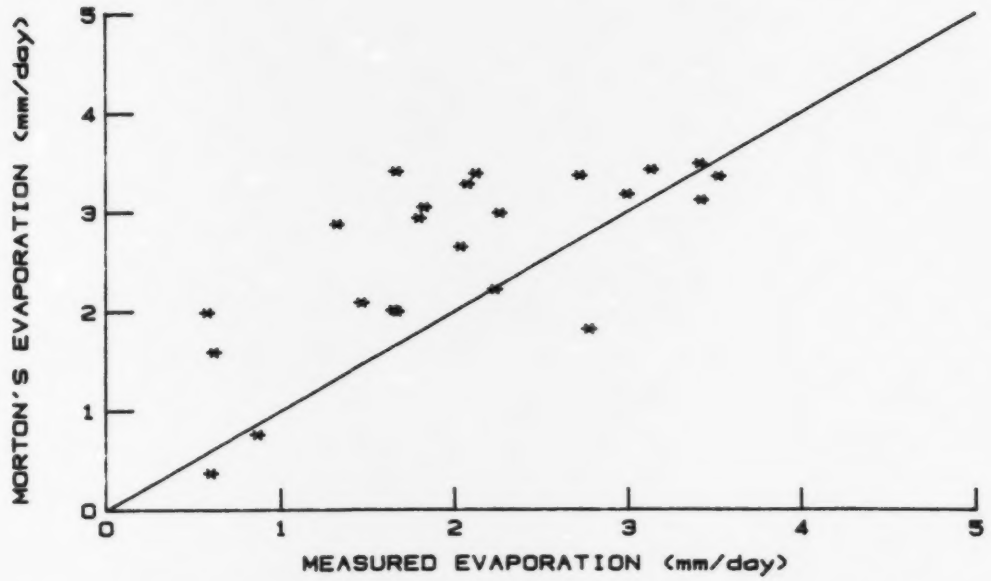

Figure 6.19 Regional estimates by CRAE (without a wind correction) plotted against measured evapotranspiration for June, July and August at Bad Lake (1981, 1982) and Saskatoon (1982, 1983).

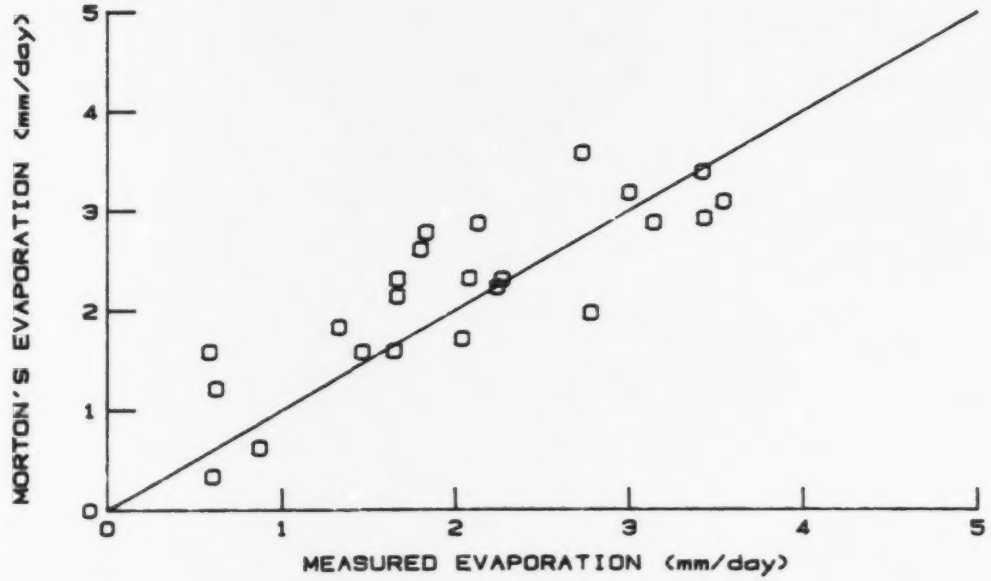

Figure 6.20 Regional estimates of evapotranspiration by CRAE (with wind correction) plotted against measured evapotranspiration for June, July and August at Bad Lake (1981, 1982) and Saskatoon (1982, 1983).

7. SUMMARY AND CONCLUSIONS.

Morton's (1983) CRAE model is based on a relationship developed by Bouchet (1963) who demonstrated that there is a negative or complementary response of potential evapotranspiration to changes in the availability of water for areal evapotranspiration. If the relationship is truly complementary it can be represented by:

$$E_T + E_{TP} = 2E_{TW}, \qquad [7.1]$$

in which E_T = actual evapotranspiration, E_{TP} = potential evapotranspiration and E_{TW} = wet-environment evapotranspiration. Morton suggests that the complementary relationship cannot be verified with currently available theoretical knowledge and empirical evidence. However, he assures the validity of Bouchet's relationship by using the complementary relationship itself as the basis for the definition of the wet-environment evapotranspiration rate; and establishes the wet-environment evapotranspiration rate equal to one-half the dry-environment potential evapotranspiration rate (as he has defined it).

Morton (1983) stresses the applicability of the CRAE model to a variety of environments:

> "Although most of the associated empiricisms were selected from the literature or were developed independently from published data, it was necessary to use a certain amount of judgement in the selection process and in making minor adjustments to obtain greater generality (e.g., in using atmospheric pressure to estimate atmospheric radiation).....This methodology has required that a correction made to obtain agreement between model and river-basin water-budget estimates in one environment must be applicable without modification in all other environments."

However, in achieving this objective, the approaches used (empirical relationships obtained from monthly data; approximate solutions for albedo; neglect of the effects of wind speed and surface roughness; calibrations against large-scale river basin water budget estimates) are such that the model loses its predictive capabilities for short-term simulations of special case environments, such as for snow cover or for single, field-sized areas. Any model, with components based on semi-theoretical considerations, and designed

to give estimates of parameters over an integrated spatial and temporal scale should not be expected to give good estimates over short time intervals under site-specific physiographical and climatic conditions.

The three major weaknesses of the CRAE model are its' radiation algorithm, the lack of a routine for estimating the net energy available for evaporation/evapotranspiration during the period of soil thawing and the assumption that the vapor transfer coefficient is independent of wind speed.

(1) Although the model is not particularly sensitive to errors in insolation data, this does not reflect the model's sensitivity to errors in the radiation estimate. Any model, based on energy balance considerations, requires that the radiation algorithms used provide accurate and reliable estimates of the net all-wave flux at the surface. The radiation model which Morton uses in CRAE and the monthly structure of the calculations result in significant errors when shorter time periods are used. It is demonstrated that there are also significant errors in the estimation of the transmittancy of clouds to extra-terrestrial radiation, the albedo and the net long-wave radiation.

(2) The assumption that the soil heat flux and storage terms are negligible restricts the use of CRAE to those periods when soil is not freezing or thawing. In the case of soil thawing an appreciable amount of available energy may be apportioned to the soil heat flux term which supplies the energy required for the phase change and soil warming. It is demonstrated that CRAE consistently overestimates actual evapotranspiration during these periods.

(3) The assumption that the vapor transfer coefficient is independent of wind speed is not appropriate. It is demonstrated that neglecting the effects of wind speed can result in significant errors, even when monthly periods are used. As well, neglecting the effect of wind speed means that the effects of surface roughness are also ignored; which limits estimates to a "Regional" scale and prevents the model from simulating the effects on evapotranspiration of natural or man-made changes to the environment.

CRAE requires accurate humidity data and the accuracies generally associated with the most commonly-used methods of humidity measurement (lithium chloride dewcell, ∓3% RH; hygrothermograph, ∓5%) are inadequate.

The model appears to provide better estimates of the "Regional" evapotranspiration during periods of high evapotranspiration demand. Possibly this is due to the fact that during these periods the differences between the actual, potential and wet-environment evapotranspiration rates and the differences between the actual rates on bare and vegetative areas are also smaller. CRAE also appears to be sensitive to abrupt changes in dominant weather within the interval of calculation, thus estimates of evapotranspiration may be affected by the length of the averaging period. However, for periods when the dominant weather remains reasonably stable, the length of the averaging period has little effect on the calculated amount and the correction procedure suggested by Morton (1983, Section C.2.5) for daily estimates is not needed.

8. RECOMMENDATIONS

It is considered that the CRAE model persented by Morton (1983) is an appropriate framework for the development of a system which is capable of providing short-term (daily) estimates of "areal" evaporation/evapotranspiration from field-sized areas having specific soil, surface cover and climatic conditions. In order to achieve this objective it is recommended that the "existing" version of the model (Morton, 1983) be modified as follows:

(1) Net All-wave Radiation - modifications and improvements to the routine which will allow:
   (a) Proper accounting for the transmittancy of clouds to extra-terrestrial radiation or the estimation of direct and diffuse components of the incident global flux.
   (b) Inclusion of site-specific albedo values and replacement of the procedure for estimating snow surface albedos.
   (c) Proper evaluation of the net long-wave flux, in particular as regards to the downward-directed component and the effects of ambient water vapor pressure on the net flux.
   (d) Estimation of components for short time periods, that is, replacement of the monthly structure with a more flexible framework.

(2) Incorporation of an algorithm to account for the effects of the soil heat used for the phase change and warming in frozen soils on the net energy available for evaporation/evapotranspiration.

(3) Incorporation of an algorithm to account for the effects of wind speed on the vapor transfer coefficient. For monthly calculations of "Regional" evapotranspiration a correction based on the deviations from the long-term mean wind speed at a site may prove adequate.

It is suggested that these modifications will improve the performance of the model for estimating evapotranspiration over integrated spatial and temporal scales.

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